STRATAL PATTERNS OF THE WILLIAMS FORK (HUNTER CANYON) FORMATION, PICEANCE BASIN, COLORADO

BY

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ABSTRACT

The Cretaceous Williams Fork Formation contains stratal packages that represent the final stage of deposition within the Cretaceous Western Interior foreland basin (Colorado). These stratal packages record changes in accommodation through time and are interpreted using a sequence-stratigraphic approach. The formation is divided into three units based on changes in stratal packaging and accommodation style. Unit 1 contains aggradationally stacked parasequences interpreted to represent deposition during highstand systems tracts. Units 2 and 3 both contain stacked, repetitive stratal packages interpreted as depositional sequences. Depositional sequences contain lowstand, transgressive and highstand systems tracts. A complicated interplay between eustatic oscillations and tectonic subsidence controls deposition of the stratal packages and depositional sequences. Eustatic oscillations control deposition of depositional sequences while tectonic subsidence forms the accommodation necessary to preserve multiple stacked stratal packages. The overall change in accommodation style from Units 1- 3 records a decrease in accommodation through time interpreted to represent a decrease in the rate of tectonic subsidence.

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CHAPTER ONE
Introduction

The purpose of this study is to describe stratal packages present within the Cretaceous Western Interior foreland basin, to interpret these packages using a sequence-stratigraphic approach and to discuss the mechanisms responsible for forming these packages. Stratal packages record changes in accommodation through time. Describing and interpreting these changes in accommodation allows for a better understanding of how a foreland basin is filled through time and what mechanisms are responsible for changes in the style of accommodation.

The Williams Fork Formation contains stratal packages and was chosen for detailed study because of extensive exposures and a noticeable vertical change in lithology and sandstone-to-shale (net-to-gross) ratios. Ages for deposition of the Williams Fork Formation are uncertain: interpretation of palynologic data places the base of the formation in the very latest Campanian to the early Maastrichtian and the top of the formation deposition at the latest Maastrichtian (Johnson and May, 1980). Interpretations by Johnson and May (1980) indicate that deposition of the formation did not continue after the end of the Cretaceous. There are no other dating techniques that have been applied to the formation to help narrow down the time constraints. If the longest amount of time interpreted from the palynological date is ~14m.y. from ~70 Ma to ~66 Ma. The shortest amount of time for deposition of the formation interpreted from the palynological data is ~8 m.y. from ~74 Ma to ~66 Ma. The formation represents the final stage of deposition within the foreland basin. Outcrops of the Williams Fork Formation are well exposed in the Laramide Piceance basin of western Colorado. This study focuses on outcrops approximately twenty miles east of Grand Junction Colorado, (1) in Coal Canyon, (2) along highway 64 between I-70 and DeBeque Cutoff Road and (3) along I-70 between Palisade and DeBeque. In the study area, the Williams Fork Formation is approximately 500 m thick.

One composite stratigraphic section and several incomplete, though more detailed, stratigraphic sections were measured and described. Parts of the composite section were correlated using the top surfaces of laterally continuous sandstones. Each portion of the composite section was measured vertically, when possible, in order to keep the section unbiased. In some areas, however, safety reasons made it impossible to continue vertically up the section and measurements would have to be taken laterally. Sandstones in the formation change thicknesses laterally even when they are laterally continuous. Because of this, and the attempt to keep the composite section unbiased, the composite section may not always appear to correspond with descriptions within the paper. For example, stacked, laterally continuous sandstones in several areas are described as becoming progressively thinner with each vertically stacked sandstone. While this is true, the composite section may have been measured in an area where a lower sandstone was at its thinnest and an upper sandstone was at its thickest resulting in the appearance that sandstones thicken vertically (Fig. 1). When this occurred, observations of lateral deposits were made using photos and binoculars. Often data such as grain size could be collected from adjacent areas even when the section could not be safely measured. The patterns described in the paper are based not only on the composite section but also on observations made lateral to the measured section. The fluvial sandstones tend to thin and thicken and therefore a sandstone drawn as fine grained and one meter thick in the composite section may in fact thicken to four or five meters and coarsen in the majority of the lateral exposures.

Data collected includes facies descriptions, sandstone body geometries and sand to mud ratios. In gerneral, grain size was measured every meter to meter and half using a grain size card and a hand lens. Grain size was always measured at the base and top of a sandstone. Grain size of sandstones was also measured in outcrops lateral to the composite section to check for lateral variations in grain size. Sandstone body geometries were recorded using binoculars and photos or by making several thickness measurements lateral to eachother within the same sandstone unit. Units were traced laterally by following the top surface of prominent sandstone units.

The Williams Fork Formation consists of two scales of stratal packaging. The larger scale divides the Williams Fork Formation into three units based on changes in the style of accommodation. The smaller-scale stratal packaging consists of repetitive stratal patterns present in Units 2 and 3 but absent in Unit 1. Chapter Two focuses on the description and sequence-stratigraphic interpretation of these stratal packages. The interpretation of the stratal packages indicates complicated changes in accommodation throughout deposition of the Williams Fork Formation. Chapter Three discusses the mechanisms acting within the foreland basin that may be responsible for these changes in accommodation style. The Appendix contains the composite measured stratigraphic section.

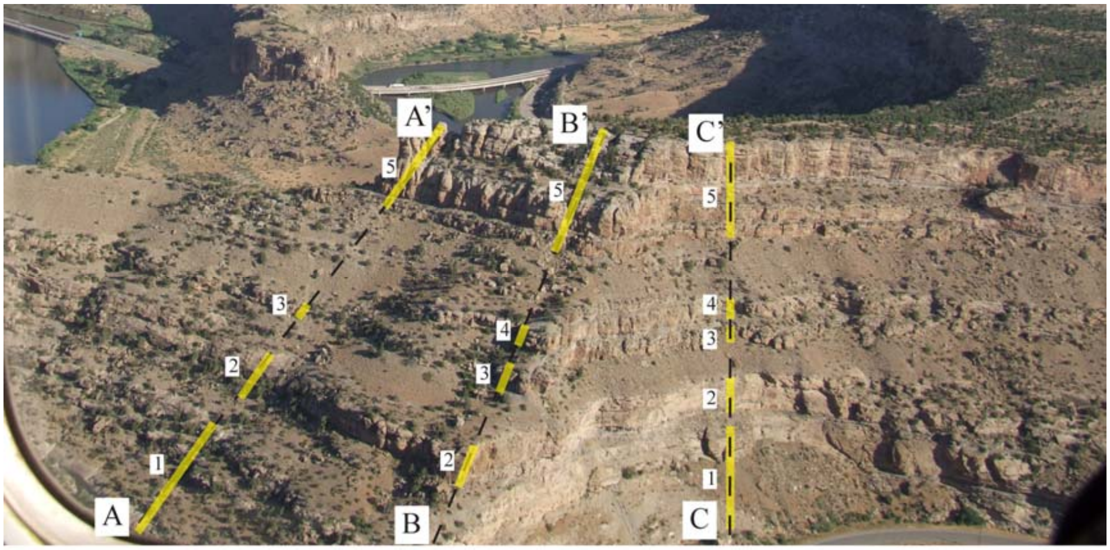

Figure 1: Three hypothetical vertical transect lines through a portion of the Williams Fork Formation showing how the measured section will look different depending on where the section is measured. The yellow lines show sandstone thickness along the transects. Each laterally continuous sandstone is numbered in each transect. If the section was measured along A-A' then it would appear as if only 4 sandstones were present because sandstone 4 pinches out. If the transect were measured along B-B' then sandstones 2, 3 and 4 would appear as if they thin progressively. Since line B-B' goes through the thickest parts of each sandstone it is the most accurate in terms of sandstone thickness. If the transect were measured along C-C', however, it would appear as if sandstone 4 were thicker than sandstone 3 because the line goes through a thicker portion of sandstone 4 and a thinner portion of sandstone 3. This is why it was important to observe lateral to where the transect was measured in order to have the most detailed description of the formation.

The Williams Fork Formation was named and described by Hancock (1925) and then eventually traced from the Grand Hogback to the Colorado-Utah state line by Fisher et al. (1960) and Collins (1976). The Williams Fork Formation is a gas producing formation (Cumella and Ostby, 2003), and previous study has focused on the reservoir architecture of the formation (Lorenz et al., 1985; Hettinger and Kirschbaum, 2003; Cole and Cumella, 2005; Pranter et al., 2007). Recently, several studies have employed high-resolution aerial light detection and ranging (LIDAR) data in order to examine sandstone interconnectivity in the lower Williams Fork Formation and to perform reservoir characterization and modeling of a point bar in the same stratigraphic interval (Pranter et al., 2007; Pranter et al., 2009).

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CHAPTER TWO

Sequence-stratigraphic interpretation of stratal patterns within the Williams Fork Formation, Piceance Basin, Colorado

ABSTRACT

The Upper Cretaceous Williams Fork Formation contains distinct stratal packages that record the final stage of fluvial fill within the Cretaceous Western Interior foreland basin in Colorado. The stratal packages reflect changes in accommodation through time and are interpreted using a sequence-stratigraphic approach. The formation is divided into three units based on facies and stratal patterns. Unit 1 contains isolated channel-fill sandstones encased in overbank deposits. The channel-fill sandstones are aggradationally stacked and represent high accommodation fill deposited during one or multiple highstand systems tracts. Units 2 and 3 also contain channel-fill sandstones and overbank deposits, however, both are composed of repetitive stratal packages that are vertically stacked. Each stratal package displays an overall upward increase in accommodation and a decrease in fluvial energy reflecting an overall rise in base level. Each stratal package is interpreted to be a depositional sequence. In Unit 2, depositional sequences consist of a basal sequence boundary overlain by a lowstand, transgressive and highstand systems tract. Depositional sequences within Unit 3 consist of a basal sequence boundary overlain by a transgressive systems tract and a highstand systems tract.

INTRODUCTION

Stratal successions record depositional events and changes in accommodation during fill of sedimentary basins. Some of these successions appear random while others display specific repetitive stratal patterns. Stratal patterns are well developed in the late Cretaceous Williams Fork Formation where they record the final stage of fill in the Cretaceous Western Interior foreland basin. Interpretation of the stratal patterns is possible using a sequence-stratigraphic approach. Sequence stratigraphy is commonly used to interpret stratal patterns in marine deposits within foreland basins (Weimer, 1983, 1984; Mitchum and Van Wagoner, 1991; Taylor, 1991; Kamola and Huntoon, 1995; Krystinik and DeJarnett, 1995; O'Byrne and Flint, 1995; Schwans, 1995; Van Wagoner, 1995; Legarreta and Uliana, 1996). The analysis of stratal patterns is not limited to foreland basins; sequence stratigraphy is also used to interpret stratal patterns along continental margins (Vail et al., 1977; Posamentier and Vail, 1988; Van Wagoner et al., 1990; Hentz and Zeng, 2003). A sequence-stratigraphic approach is used to analyze stratal patterns in Miocene marine strata of the northern margin of the Gulf of Mexico at a sequence set scale (Hentz and Zeng, 2003).

Stratal patterns are recognizable in fluvial deposits (Fouch et al., 1983; Wright and Marriott, 1993; Olsen et al., 1995; Schwans, 1995; Shanley and McCabe, 1995; Van Wagoner, 1995; Catuneanu and Elango, 2000), but are not as well documented as in the marine settings. The majority of the work on fluvial stratal patterns focuses on deposits within foreland basins due to high accommodation rates, which allow for the preservation of thick fluvial successions. Stratal patterns in fluvial strata are documented in a normal fault-bounded (rift) basin in the Lower Jurassic Statfjord Formation in the North Sea (Van Wagoner et al., 1995).

FLUVIAL RESPONE TO BASE-LEVEL CHANGE

Sequence stratigraphy was originally applied to marine deposits where changes in base level are identified by interpreting changes in stacking patterns of the marine deposits (Vail et al., 1977). This is possible because marine systems respond to changes in base level in predictable ways (Vail et al., 1977, 1984). Further study revealed that fluvial systems also respond to changes in base level in predictable ways, in order to maintain an equilibrium profile and, therefore, can also be analyzed using sequence stratigraphy (Posamentier and Vail, 1988; Shanley and McCabe, 1994). Fluvial systems respond to changes in base level through aggradation or incision of the fluvial profile as well as through changes to the morphology of the fluvial system (Richards, 1996; Blum and Törnqvist, 2000). In fluvial systems base level refers to an "…equilibrium surface…above which a particle can not come to rest and below which deposition and burial is possible" (Sloss, 1962, p. 1051). Above this surface sediment cannot be preserved and below this point sediment can be preserved. The position of this surface can change through time. In general, changes in base level result in changes to the gradient and morphology of the fluvial system.

A drop in base level can lead to an increased gradient in fluvial systems resulting in incision (Miall, 1996). The incision occurs as a response to an increase in the power, competency and capacity of the fluvial system, and an increase in the potential for sediment bypass and erosion (Richards, 1996). If the increase in fluvial gradient is low, changes in the morphology, sinuosity and load characteristics may be sufficient to establish a new equilibrium profile. If the increase in fluvial gradient is large, incision will occur, forming an incised valley in the lower reaches of the fluvial system (Van Wagoner et al., 1990). Incision occurs through the process of knickpoint migration where the knickpoint is the point along the longitudinal fluvial profile where the gradient increases (Schumm 1993). Incision initiates at the knickpoint and continues upstream from that point. Incision may be coupled with other changes in the fluvial system such as a widening of the fluvial channel, the formation of incised valleys that are much wider than the rivers within them, a relative decrease in channel sinuosity and a change to coarser, bed-load dominated systems (Schumm, 1981; Marzo et al., 1988; Van Wagoner et al., 1990). These fluvial systems will continue to incise until a new equilibrium is reached. Commonly, fluvial systems do not have time to reach equilibrium if a base-level rise quickly follows the base-level fall (Richards, 1996).

Base-level rise can decrease the gradient of a fluvial system and result in aggradation (Richards, 1996). A base-level rise results in a lower gradient fluvial system and can result in the reduction of stream power, discharge, and the ability of the fluvial system to maintain its sediment load (Schumm, 1993). This results in deposition of the sediment load and aggradation of the fluvial system. Channels tend to become less bed-load dominated and channel sinuosity increases relative to the higher gradient fluvial system (Richards, 1996).

Once a fluvial system achieves equilibrium conditions, it will not aggrade or incise. Equilibrium can be reached if base level remains constant for an extended period of time (Schumm, 1993). During times of equilibrium conditions, sinuous channels may rework lateral deposits through point bar migration and channel meandering (Schumm, 1993). Flood events cause active meandering channels to build up their levees and surrounding floodplain areas, which result in changes to the slope of the surrounding floodplain. During a flooding event, a channel may break its levee and begin to follow a new course. This new course will follow the highest gradient through the floodplain (Richards, 1996). Braided systems also rework lateral deposits through avulsion, channel switching and the lateral migration of braid bars (Miall, 1996).

RECOGNIZING SYSTEMS TRACTS IN FLUVIAL SUCCESSIONS

Systems tracts are a linkage of contemporaneous depositional systems (Brown and Fisher, 1977) that can be defined by their bounding surfaces, their position within the sequence and their parasequence stacking pattern (Van Wagoner et al., 1988). Three systems tracts recognized in the marine realm are the lowstand systems tract, the transgressive systems tract and the highstand systems tract (Van Wagoner et al., 1988). In the marine realm, the lowstand systems tract is bounded below by a sequence boundary and above by the first widespread, major transgressive surface (Posamentier and Vail, 1988; Van Wagoner et al., 1988). The transgressive systems tract is bounded below by the first widespread, major transgressive surface and above by the maximum flooding surface (Posamentier and Vail, 1988; Van Wagoner et al., 1988). The highstand systems tract is bounded below by the maximum flooding surface and above by a sequence boundary (Posamentier and Vail, 1988; Van Wagoner et al., 1988). Surfaces that bound the systems tracts within marine deposits, such as the major transgressive surface and maximum flooding surface, are difficult to identify within fluvial deposits. Systems tracts within fluvial deposits are identified based on changes in stacking patterns, changes in fluvial facies, and position within the sequence (Shanley and McCabe, 1995; Van Wagoner, 1995; Richards, 1996). Changes in fluvial characteristics reflect changes in base level and gradient and allow changes in fluvial characteristics observed in the stratigraphic record can be used to identify systems tracts.

The lowstand systems tract is underlain by a sequence boundary, which is initiated by a base-level fall resulting in negative accommodation, erosion, incision, and sediment bypass. In fluvial systems, sequence boundaries are recognized by regional truncation of older strata and a basinward shift in facies across a boundary (Van Wagoner, 1995). The basinward shift in facies may be recognized by a change from lower energy fluvial deposits below the boundary to higher energy fluvial deposits above the boundary (Van Wagoner, 1995). The incision process results in an increase in the gradient of the fluvial system. The late lowstand is characterized by low accommodation and the onset of sediment accumulation (Van Wagoner, 1995). During this time, a slow rise in base level results in deposition within the fluvial system, and a slight decrease in gradient. In the late lowstand, sediment begins to accumulate in the fluvial system as base level begins to rise, promoting reworking of lateral deposits and resulting in deposition of multi-story and interconnected channel-fill sandstones (Van Wagoner, 1995; Richards, 1996). Multi-story, stacked channel-fill sandstones, deposited by braided or meandering channels, typically dominate late lowstand deposits. Multi-story, stacked point bar deposits, described in the Lower Jurassic Statfjord Formation of the North Sea, are bounded below by a sequence boundary and are interpreted to represent the lowstand systems tract (Van Wagoner et al., 1995). Lowstand systems tract deposits typically represent the highest energy fluvial deposits within a depositional sequence (Richards, 1996). In this study, the terms high and low-energy fluvial systems are relative terms. The designation of a fluvial system as high energy denotes that the fluvial system has greater competency, capacity and flow velocity relative to low-energy fluvial systems. This is interpreted from the rock record by coarser sediment and bedforms indicative of higher flow velocities than in low energy fluvial systems.

The transgressive systems tract results from increased accommodation and aggradation of the fluvial system as the rate of base-level rise rapidly increases (Van Wagoner, 1995). The rise in base level results in successively lower gradients and fluvial systems become less competent and energetic (i.e. straight channels become more sinuous, discharge decreases) (Richards, 1996). The increase in accommodation also allows for the limited preservation of overbank fines because there is less time for the lateral migration of channels. Preservation of overbank fines is greater in the transgressive systems tract than in the late lowstand systems tract. Fluvial deposits of the transgressive systems tract are typically characterized by single-story channel-fill deposits interbedded with overbank fines (Richards, 1996). The top of each channel-fill represents a parasequence boundary and the parasequences may display a backstepping stacking pattern (Richards, 1996). Fluvial deposits within transgressive systems tracts in both the Statfjord Formation of the North Sea and in the Castlegate Formation of Utah occur as single-story channel-fill sandstones interbedded with overbank mudstones and siltstones (Van Wagoner et al., 1995).

The highstand systems tract is characterized by a gradual decrease in the rate of base-level rise. The continued rise in base level results in the lowest fluvial gradients within the entire sequence. Fluvial systems typically are fine-grained, meandering systems, and the resulting deposits have a low net-to-gross ratio (Van Wagoner, 1995). Thick overbank deposits interrupted by isolated, fine-grained, channel-fill deposits characterize highstand systems tract deposits. Channel-fill sandstones are less laterally continuous than the channel-fill sandstones in the lowstand or transgressive systems tracts. The overbank deposits are typically red to green if the alluvial plain is well drained but can be brown to dark brown with coal and lake deposits if the alluvial plain is poorly drained (Wright and Marriott, 1993; Van Wagoner, 1995). Parasequence development during the highstand systems tract can occur quickly or slowly. If parasequences develop quickly, they are characterized by a high accommodation to sediment supply ratio (Richards, 1996). This results in poorly connected, vertically isolated channel-fill deposits that are encased in overbank deposits (Richards, 1996). High accommodation allows little time for lateral reworking resulting in the preservation of overbank fines. If parasequences develop slowly, they are characterized by a low accommodation to sediment supply ratio, which results in time for lateral reworking and an increase in channel-fill connectivity (Richards, 1996). The Turonian to Santonian strata of the Kaiparowits Plateau contain approximately 125 m of isolated, lens shaped, channel-fill sandstones encased in overbank fines and are interpreted as highstand deposits (Shanley and McCabe, 1995). The isolated nature of the channel-fill sandstones and the low net-to-gross ratio indicate that parasequences developed quickly with little time for lateral migration of the channels.

Changes in sediment supply can also produce changes in fluvial style and fluvial architecture. In this study, however, sediment supply is assumed to be relatively constant throughout deposition of the Williams Fork Formation. Changes in base level, not changes in sediment supply, are considered the primary mechanism responsible for the fluvial stacking patterns within the Williams Fork Formation. There are several reasons that sediment supply is assumed to be constant in this study. First, during development of the Western Interior foreland basin the orogenic belt is a source for sediment. Secondly, the basin is filled (Decelles, 2004) allowing sediment from the orogenic belt to be readily transported to distal portions of the basin. Thirdly, during the late Cretaceous the climate in the study area was warm and humid with high precipitation year round (Wolfe, 1979; Golovneva, 2000). In this climate extensive erosion would have occurred delivering high amounts of sediment to the basin. Several other similar studies located within the Western Interior foreland basin have also attributed changes in fluvial architecture to changes in base level and assumed that sediment supply is constant and not a primary control (Shanley and McCabe, 1995; VanWagoner, 1995).

REGIONAL STRATIGRAPHY

The Williams Fork Formation is exposed discontinuously from the Colorado-Utah state line east to the Grand Hogback near Rifle, Colorado (Fisher et al., 1960; Collins, 1976) (Fig. 1). The Williams Fork Formation overlies the marine Rollins Sandstone Member of the Iles (Mt. Garfield) Formation and is capped by the Paleocene to Eocene fluvial and lacustrine Wasatch Formation. The exact timing of deposition of the Williams Fork Formation is debated (Johnson and May, 1980; Patterson et al., 2003) (Fig. 2). Ages for deposition of the Williams Fork Formation are uncertain: interpretation of palynologic data places the base of the formation in the latest Campanian to the early Maastrichtian and the top of the formation at the latest Maastrichtian (Johnson and May, 1980). Interpretations by Johnson and May (1980) indicate that deposition of the formation did not continue after the end of the Cretaceous. There are no other dating techniques that have been applied to the formation to help narrow down the time constraints. The longest amount of time interpreted from the palynological date is ~14m.y. from ~70 Ma to ~66 Ma. The shortest amount of time for deposition of the formation interpreted from the palynological data is ~8 m.y.

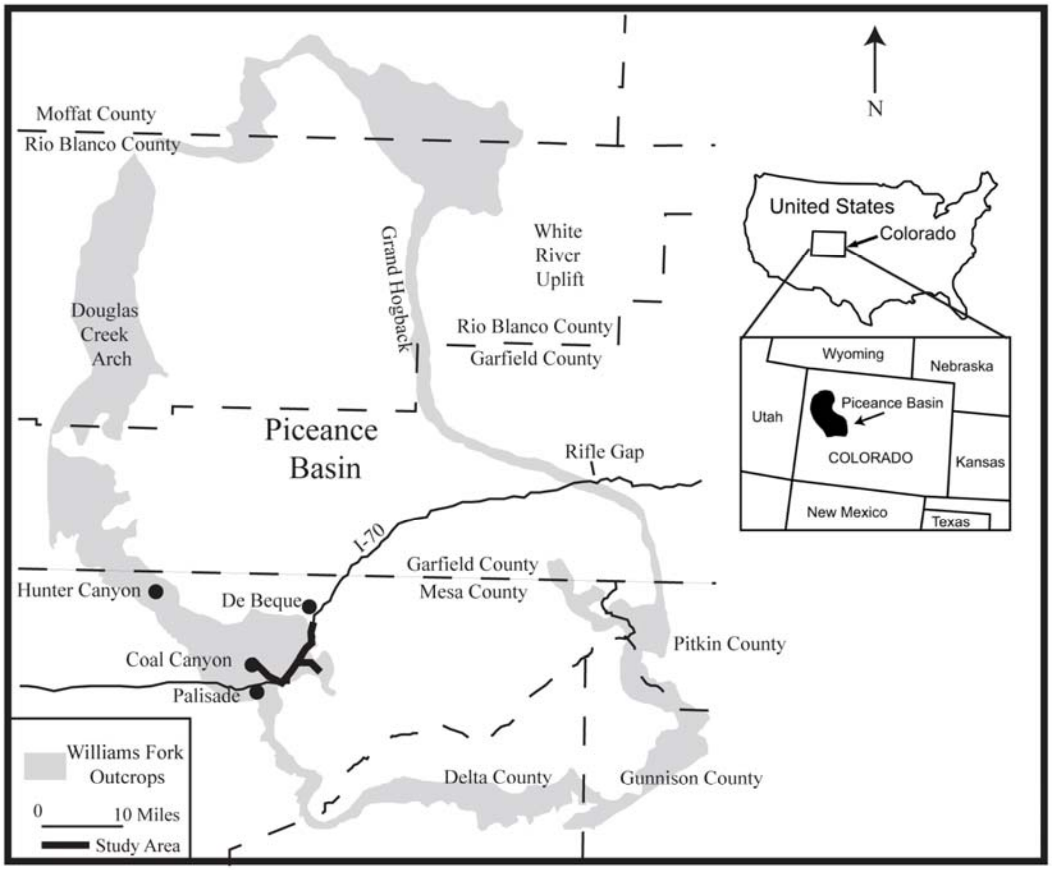


Figure 1: Map showing location of the Williams Fork Formation outcrops within the Piceance Basin (Colorado) and location of this study (Modified from Pranter et al., 2007).

from ~74 Ma to ~66 Ma. The Williams Fork Formation and its stratigraphic equivalents can be traced throughout the Piceance basin, northward into the Sand Wash Basin, and westward into eastern Utah (Erdman, 1934; Warner, 1964; Collins, 1976; Madden, 1989; Roehler, 1990) (Fig. 2). A nomenclatural change occurs as the formation is traced west from the study area into the Book Cliffs of Western Colorado (Fig. 2).

This study focuses on outcrops of the Williams Fork Formation located in Coal Canyon, along highway 65 between I-70 and DeBeque Cutoff Road, and along I-70 between Palisade and DeBeque, Colorado (Fig. 1). One complete composite stratigraphic section and several incomplete, but more detailed, stratigraphic sections were measured. Data collected includes facies descriptions, sandstone body geometries and net-to-gross ratios. Sections were correlated by walking out the upper surfaces of laterally continuous channel-fill sandstones. These data are the basis for the interpretations made.

In the study area, the formation can be divided informally into a lower sand poor interval (basal ~150 to 200 m) and upper sand rich interval (upper ~350 m) (Fisher et al., 1960) (Fig 3). The lowest ~15-75 m of the formation is coal rich and referred to as the Cameo-Wheeler coal zone (Hettinger and Kirschbaum, 2003). A white, kaolinized, quartz pebble conglomerate occurs at the top of the Williams Fork Formation and is informally called the Ohio Creek conglomerate (Johnson and May, 1980). The Ohio Creek conglomerate is ~15 m thick in the study area and stratigraphically equivalent to the Ohio Creek Member of the Hunter Canyon and Mesa Verde Formations (Johnson and May, 1980). Both the Cameo-Wheeler coal zone and Ohio Creek Conglomerate are informal divisions within the study area and therefore the nomenclature is not used within this study.

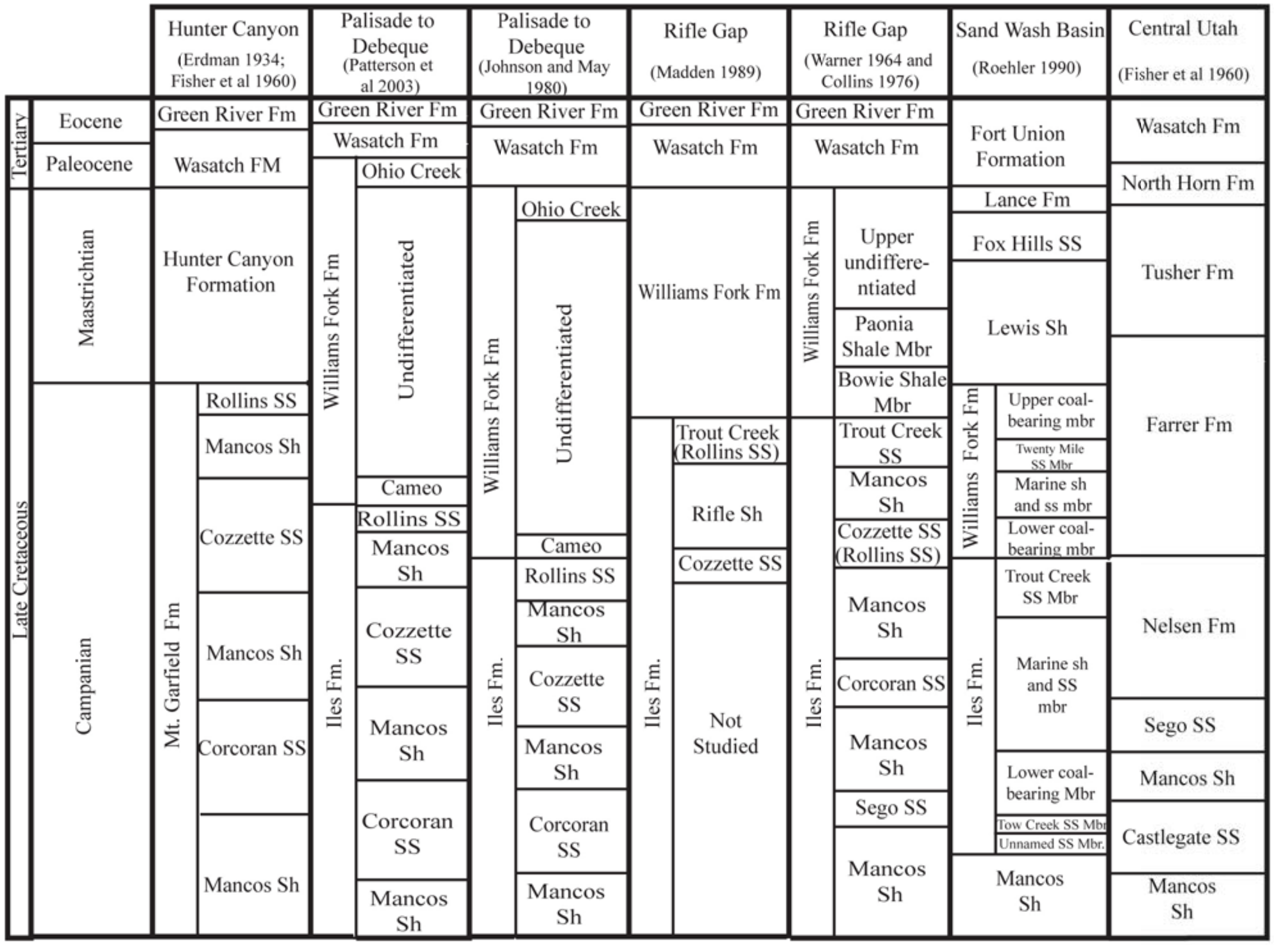


Figure 2: Stratigraphic sections from Hunter Canyon (CO), Palisade to Debeque (CO), Rifle Gap, Sand Wash Basin (CO and WY) and Central Utah showing changes in stratigraphic nomenclature between localities and disputes in timing of deposition. The Hunter Canyon stratigraphic section displays the stratigraphic nomenclature and timing used in this study.

In general the Williams Fork Formation and equivalent strata coarsen toward the orogenic belt in central Utah and fine toward the Grand Hogback in central Colorado. The average grain size of sandstones in the stratigraphic equivalent Farrer and Tuscher Formations of Utah ranges from fine to medium (Farrer Formation) to coarse grained (Tuscher Formation), while grain sizes in the Williams Fork Formation of Coal Canyon and the Grand Hogback area range from fine-lower at the base to a maximum of medium grain size (Fisher et al., 1960; Johnson and May, 1980; Madden, 1989). Grain size trends and paleoflow indicators suggest that the fluvial deposits of the Williams Fork Formation are part of a generally west to east transport system (Fisher et al., 1960) within the Western Interior foreland basin.

Figure 3: Panorama of the Williams Fork Formation looking northwest of I-70, northeast of Cameo, Colorado showing the contrast between the sand poor interval and the sand rich interval. This is not the complete succession of the Williams Fork Formation; it shows the upper 90 m of the sand poor interval and lowest ~100 m of the sand rich interval.

STRATIGRAPHY

In this study, the Williams Fork Formation is divided into three main units. The basal ~130 meters (Unit 1) is a slope forming unit and is dominated by fine grained deposits, coal seams, and isolated, lens shaped sandstones. Units 2 and 3 are both cliff-forming units with laterally continuous sandstone cliffs that extend up to 10 km and are up to 80 m in thickness. The middle ~320 m (Unit 2) consists of laterally continuous sandstones and lesser amounts of mudstone. The upper ~70 m (Unit 3) consists of laterally continuous sandstones, mudstones and paleosols.

Unit 1

Unit 1 consists of isolated sandstone lenses that are encased within mudstones and coals (Fig. 4). The sandstones are single-story with a maximum thickness of seven meters and thin laterally to less than 0.5 m. The sandstone lenses have an observed lateral continuity ranging from ~40 to 800 m. Each sandstone lens overlies a scoured base and an upward-fining grain size profile ranging from fine lower to fine upper at the base to very-fine lower to fine lower at the top. Tabular and trough cross bedding is abundant at the base of each sandstone lens (Fig. 5). Cross bed thickness is ~0.2 to 0.5 m at the base of the sandstone lens and decreases vertically to ~0.1 to 0.2 m at the top. Current ripple stratification occurs at the top of the sandstone lenses (upper most 0.1 m to 1 m), overlying the trough and tabular cross stratification (Fig. 5). Lateral accretion surfaces occur in many of the isolated sandstone lenses (Fig. 6). Rip-up clasts occur locally along the basal scours of the sandstone lenses.

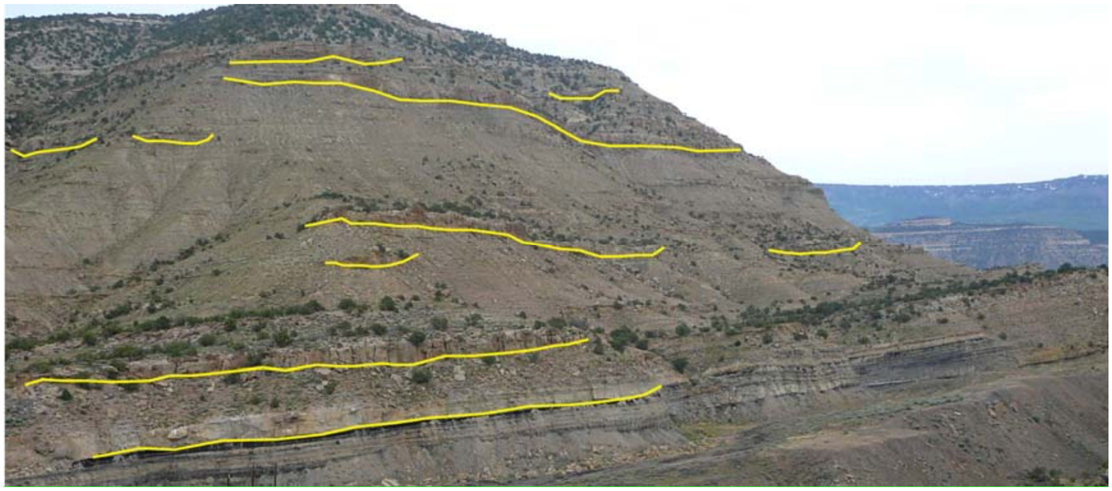


Figure 4: Lower portion of Unit 1 showing isolated channel-fill deposits encased in fine-grained deposits. Yellow lines outline the scour base of several of the more prominent channel-fills.

These sandstone lenses represent fluvial channel-fill deposits. The upward fining profile with lateral accretion surfaces is typical of deposition from point bar migration (Allen, 1963; Miall, 1996; Bridge, 2006) The presence of lateral accretion surfaces indicates that the channels had a lateral migration component (Allen, 1963). Since the majority of the channel-fill sandstones are dominated by lateral accretion surfaces the channel-fills are interpreted to be deposited by meandering channels. The development of meandering channels indicates deposition in a low gradient system such as a floodplain or coastal plain (Miall, 1996).

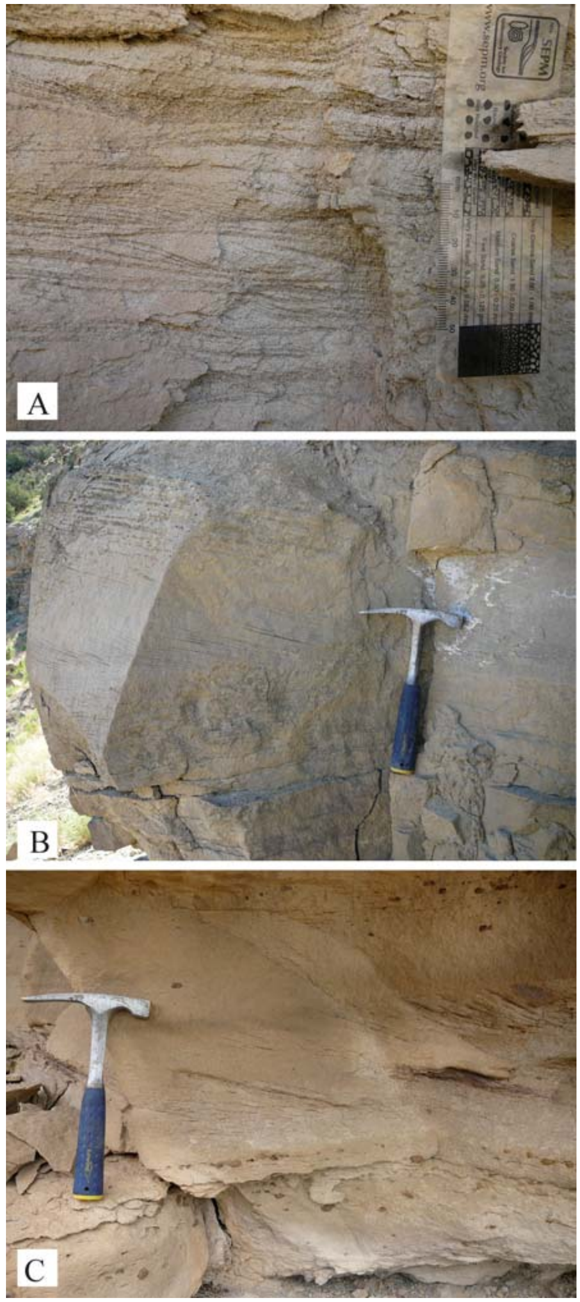


Figure 5: Several features from a channel-fill sandstone within Unit 1. A) Ripple stratification in the upper 10cm of a channel-fill deposit. B) Tabular cross stratification in the lower one meter of a channel-fill deposit. C) Tabular cross stratification an rip-up clasts near the scour base of a channel-fill deposit.

Figure 6: Lateral accretion surfaces in a channel-fill sandstone in Unit 1. Lateral accretion surfaces represent point bar migration in a meandering fluvial system.

Carbonaceous siltstones, mudstones, and rare beds of micritic limestone are interbedded with the channel-fill sandstones. The siltstones and mudstones are poorly laminated to non-laminated. The siltstones and mudstones range from dark brown to black and contain abundant plant fragments. Siltstones and mudstones represent overbank and floodplain deposits associated with the meandering channels. The dark brown to black color of the mudstones is typical of deposition in a poorly drained floodplain (Potter et al., 2005). The mudstones and siltstones appear non-laminated and featureless likely due to rooting and bioturbation, which is abundant in floodplains (Miall, 1996; Bridge, 2006). Two micritic limestone beds, 0.25 to 0.5 m thick, are traced laterally for 20 to 30 m in the lower 30 to 40 m of Unit 1 within Coal Canyon. The micritic limestone beds appear structureless (Fig. 7), are interbedded with siltstones and mudstones and are scoured out laterally by channel-fill sandstones. The micritic limestone beds are resistant, forming small ledges in the float-dominated slope. It is possible that thinner micritic limestone beds may be present in the poorly

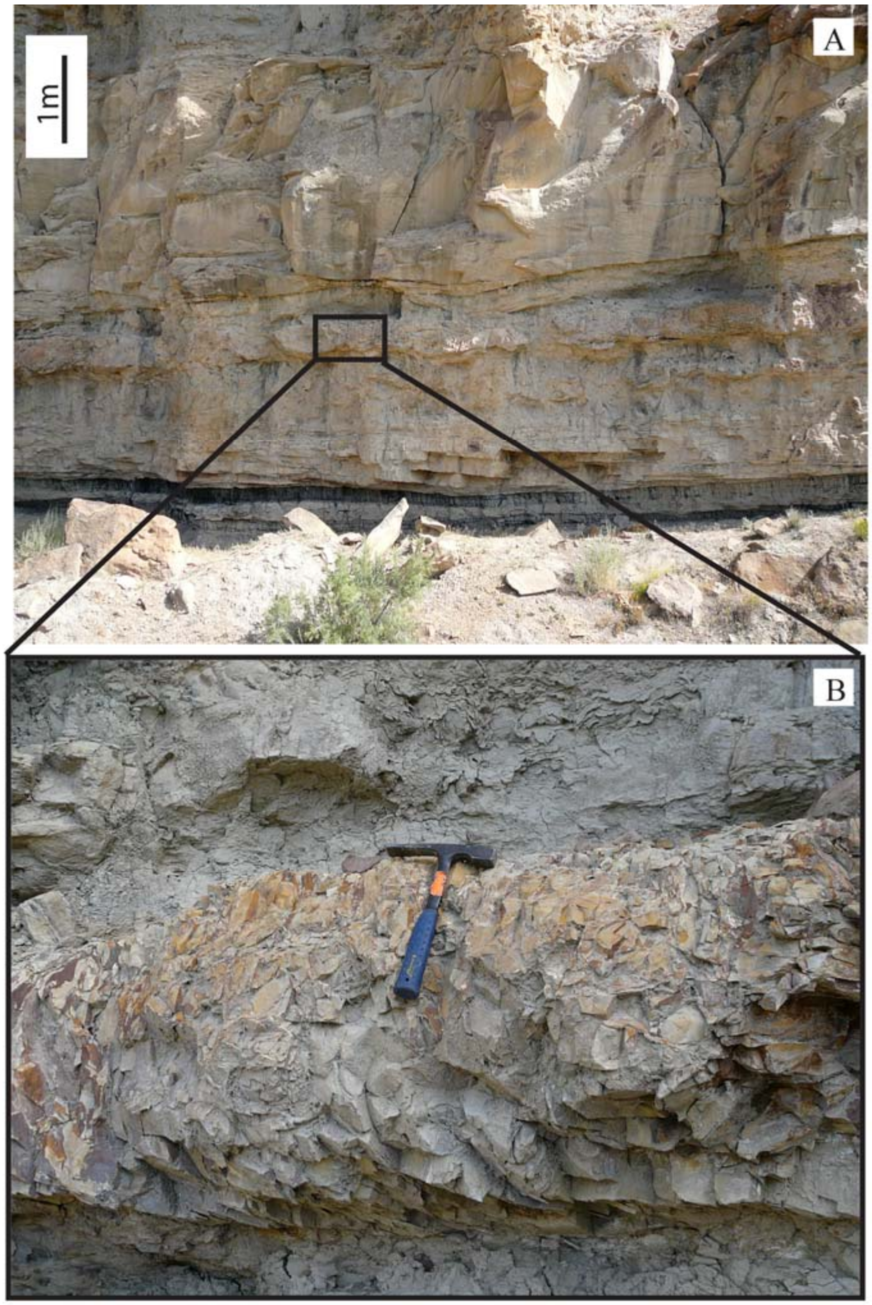


Figure 7: A) Location of one of the micritic limestone beds within Unit 1. The bed is ~.5 m thick and is a resistant layer compared to the fine grained deposits surrounding it. B) Close up of the micritic limestone bed.

exposed outcrop. The micritic limestone beds indicate low energy deposits and perhaps represent small ponds, which is consistent with the interpretation of a poorly drained floodplain.

Coal beds are limited to the basal 15 m of Unit 1 within the study area and are interbedded with organic rich siltstones, mudstones and channel-fill sandstones (Fig 8). Contacts between the coal and the organic rich siltstones and mudstones are sharp. Four distinct coal beds were measured ranging from 0.5 to 3 m in thickness. The coals are clean and not intercalated with silt. Weathering of the lower part of Unit 1 results in poor outcrop exposure of the coal beds. Well log data within 10 to 15 km north and north east of the study area suggests that in some locations more than four coal beds are present within the lower part of Unit 1, with pronounced lateral variations in thickness of the coal beds. The coals represent deposition in mires (Hancock, 1925).

Sequence Stratigraphy of Unit 1

The top of each channel-fill sandstone, and its laterally equivalent overbank deposit, is a parasequence boundary, which represents a base-level rise (Kamola and VanWagoner, 1995). The rise in base level allows for the preservation of thick overbank deposits overlying the channel-fill sandstones. Parasequence boundaries may also be present within overbank successions but are difficult to identify without the presence of a channel-fill deposit. There are at least twenty vertically stacked parasequences within Unit 1 It is likely that more parasequences are present but are not identified because the overbank expression of the parasequence boundary was not observed. There is very little variability in the vertical stacking of these parasequences. The fluvial characteristics (i.e. average grain size, cross-bed thickness, average channel-fill sandstone thickness) do not change between parasequences indicating that there is no trend in parasequence stacking (e.g. backstepping) from the base of the unit to the top. The lack of a vertical trend suggests that the parasequences are aggradationally stacked as opposed to progradationally or retrogradationally stacked, either of which would result in a vertical change in the characteristics of the channel-fills from lower

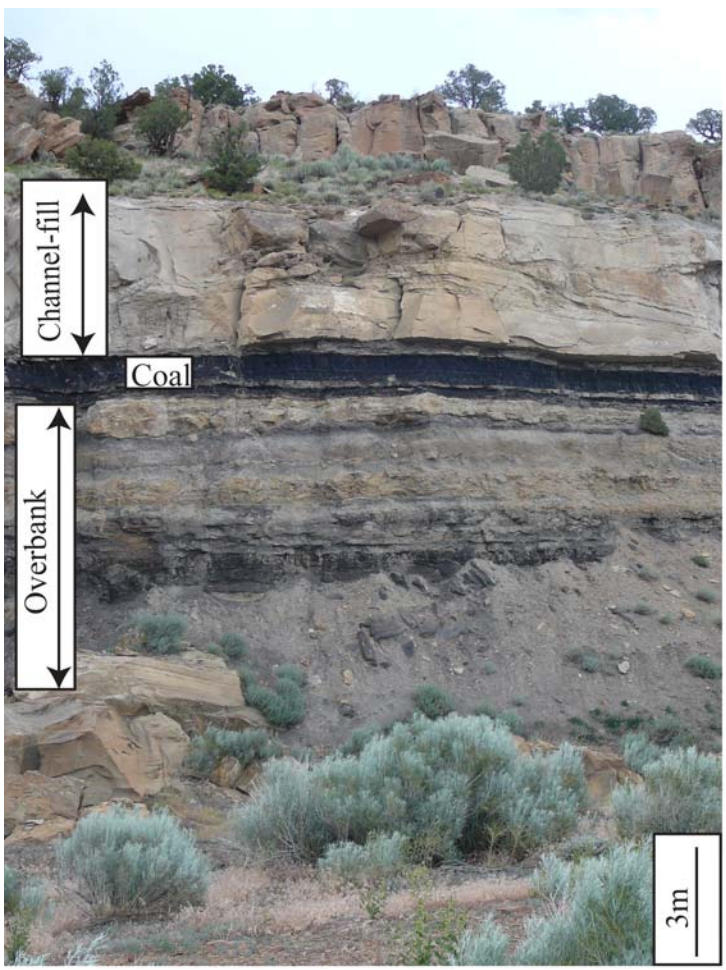


Figure 8: Coal seam and overlying channel-fill sandstone within Coal Canyon (basal 15 m of Unit 1).

energy to higher energy or higher energy to lower energy respectively (Posamentier and Vail, 1988).

The observed parasequences are well-developed with each representing the formation of accommodation and the subsequent filling of the accommodation by single-story channel-fill sandstones, with no aggradational component, and their laterally equivalent floodplain. The absence of an aggradational component to the channel-fill sandstones suggests that accommodation is formed in distinct pulses rather than continuously. A slow and continuous formation of accommodation would result in an aggrading channel system with slight lateral reworking. There is evidence of lateral reworking by point bar migration in the majority of the channel-fill sandstones. The lateral continuity of the channel-fill sandstones, however, is limited, resulting in the isolated channel-fill sandstones encased in overbank deposits. These parasequences are indicative of those that develop quickly with a high accommodation to sediment supply ratio.

Surfaces of base-level fall are not identified within Unit 1. There are no abrupt changes in fluvial morphology, increases in grain size or increases in cross bed thickness that could be interpreted to reflect a base-level fall or sequence boundary (Fig. 9) (e.g. Shanley and McCabe, 1995). It is possible that there are sequence boundaries present but they are not recognized because they occur as interfluv expressions of the sequence boundary. The low gradient fluvial channel-fill sandstones with limited lateral continuity, vertically isolated nature of the channel-fill sandstones and aggradational parasequence stacking pattern are typical of deposits within a highstand systems tract indicating that Unit 1 represents deposition during one or multiple highstand systems tracts (Richards, 1996).

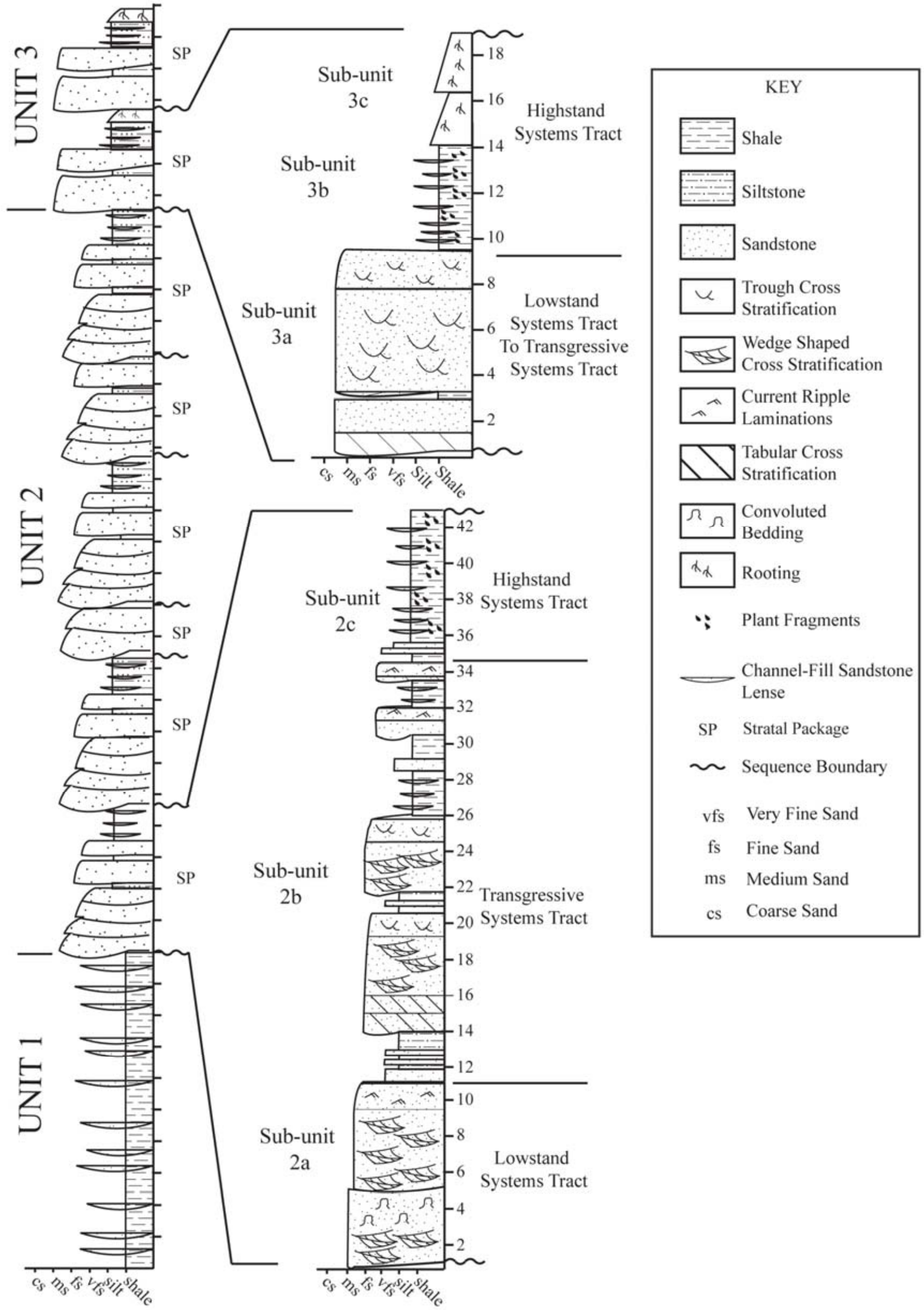
Unit 2

Unit 2 is ~310 m in thickness and is divided into multiple vertically stacked stratal packages. Each stratal package is divided into three sub-units: Sub-unit 2a (laterally continuous nested channel-fill sandstone complex), Sub-unit 2b (stacked single-story channel-fill sandstones separated by overbank fines), and Sub-unit 2c (small scale, aggradationally stacked, isolated channel-fill sandstones that are encased in overbank fines).

Sub-unit 2a

Sub-unit 2a consists of a nested sandstone complex that ranges in thickness from 8 to 20 m and maintains its thickness laterally (Fig. 10). The nested sandstone complexes can be traced laterally for the length of the exposed section (up to 10 km). Each nested sandstone complex contains multiple internal scours that represent the bases of individual sandstones that are vertically and laterally amalgamated. Each individual sandstone ranges from four to eight meters in thickness. Each sandstone displays an upward-fining profile that ranges from fine upper at the base to fine lower at the top. In some instances the upward-fining profile is not fully preserved but is truncated by an overlying scour. The average grain size of the sandstones in Sub-unit

Figure 9: (On next page) Schematic stratigraphic section of Williams Fork Formation showing Units 1 through 3 and detailed stratigraphic sections of representative stratal packages from Units 2 and 3 with sequence stratigraphic interpretation. Schematic does not show actual thickness or number of sequences present within each Unit. Vertical axis shows thickness in meters.


UNIT 3
UNIT 2
UNIT 1
SP
Sub-unit 3c
Sub-unit 3b
Sub-unit 3a
Highstand Systems Tract
Lowstand Systems Tract To Transgressive Systems Tract
Sub-unit 2c
Sub-unit 2b
Sub-unit 2a
Highstand Systems Tract
Transgressive Systems Tract
Lowstand Systems Tract
cs ms fs vfs silt shale
cs ms fs vfs Silt Shale
KEY
Shale
Siltstone
Sandstone
Trough Cross Stratification
Wedge Shaped Cross Stratification
Current Ripple Laminations
Tabular Cross Stratification
Convoluted Bedding
Rooting
Plant Fragments
Channel-Fill Sandstone Lense
SP Stratal Package
Sequence Boundary
vfs Very Fine Sand
fs Fine Sand
ms Medium Sand
cs Coarse Sand

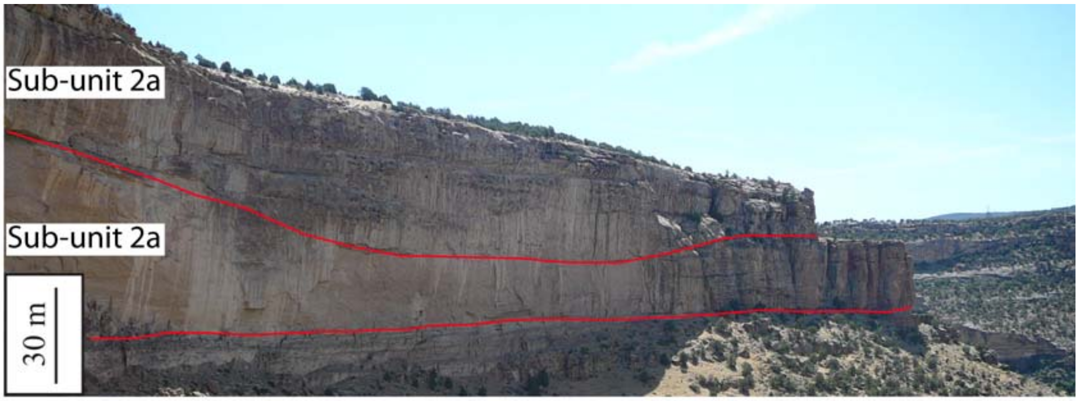


Figure 10: Panorama along I-70 showing the scale of 2 nested channel-fill sandstone complexes (two occurrences of Sub-unit 2a) in Unit 2. These particular nested channel-fill complexes can be traced laterally for 10 km.

2a is double the average grain size of the sandstones present in Unit 1 and represents the largest grain size of sandstones within Unit 2. Large-scale trough and wedge shaped cross-bed sets, ranging from 0.5 to 1.5 m in thickness, occur throughout the nested sandstone complex (Fig.11). Within an individual sandstone body the cross bed thickness decreases vertically from the scoured base to the top, but there is no vertical trend in cross bed thickness within each nested sandstone complex as a whole.Rip-up clasts are abundant along the basal scour of the complex and along internal scours. Unlike the sandstones in Unit 1, lateral accretion surfaces are absent.

Sub-unit 2a represents a complex of nested and amalgamated fluvial channel-fill deposits. The increase in grain size of the channel-fill deposits from Unit 1 to Sub-unit 2a indicates an increase in energy of the fluvial system, which is interpreted to reflect an increase in the gradient (eg. Van Wagoner et al., 1995). Large-scale cross stratification reflects deposition from the downstream migration of large-scale barforms (Mial, 1996; Bridge, 2006). The absence of lateral accretion surfaces

Figure 11: The base of Sub-unit 2a showing trough cross stratification and wedge shaped cross stratification. The dotted lines bound wedge shaped cross stratification interpreted to be deposited by the migration of a downstream migrating bedform. The top of the bedform in capped by ripples indicating flow in the same direction as the barform. The solid glack line represents the scour base of the overlying nested channel-fill deposit. Note that the thickness of the cross stratification does not display an upward thinning trend. Staff is 1.5m for scale.

indicates that the system did not have a promiment lateral migration component and was not highly meandering (Allen, 1963). Although downstream migrating bars can occur in both meandering an braided river systems (Bridge, 2006), the dominance of the downstream migrating bars without any lateral migrating point bars deposits along with the nested nature of the multiple channel-fills suggests that the system was most likely a braided river system (Miall, 1996; Bridge and Lunt, 2006). These deposits are similar to those described by Miall (1993) and Van Wagoner et al. (1995) in the Castlegate Formation. The fluvial deposits in the Castlegate formation are interpreted as braided stream deposits dominated by the downstream migration of barforms. Given enough time, braided river systems can undergo avulsion and channel switching resulting in the formation of a braid plain (Miall, 1996). The absence of overbank deposits within Sub-unit 2a indicates that enough time elapsed during deposition for the fluvial system to rework lateral deposits in a wide area.

Sub-unit 2b

Sub-unit 2b always directly overlies Sub-unit 2a. Sub-unit 2b ranges from 8 to 16 m in thickness and consists of laterally continuous sandstones interbedded with mudstones and siltstones. Within each occurrence of Sub-unit 2b, two to four laterally continuous sandstones are vertically separated from one another by 0.5 to 2 m of siltstone and mudstone deposits (Fig. 12). Although the sandstones are laterally continuous they pinch and swell and their thickness do not remain constant. Each sandstone consists of a scoured base and a single fining upward profile. Unlike Sub-unit 2a, there is no vertical amalgamation of sandstones. The thicknesses of the stacked sandstones decreases vertically throughout the sub-unit with the basal sandstone ranging from two to five meters in thickness (measured at the thickest sections of the sandstones) and the top sandstone ranging from one to two meters in thickness (measured at the thickest sections of the sandstones). Thicker sandstones extend laterally for the length of the exposure (up to six kilometers). There are two grain-size trends present within Sub-unit 2b. Each sandstone within the sub-unit displays an upward-fining sequence from fine lower to fine upper at the base to fine lower to very-fine upper at the top. The second grain size trend is an overall decrease in average grain size from the lowest sandstone in the sub-unit to the highest. Bedding is compound and complex with large-scale cross beds that are overlain by current-ripple laminations. The base of each sandstone is dominated by tabular, trough and wedge-shaped cross beds as large as 1.5 m in the thicker sandstones and 0.1 to 0.2 m in the thinner sandstones. Within each of the sandstones the cross bed thickness decreases vertically with current ripple laminations capping the sandstone (Fig. 13). Mud drapes can occur in association with the current-ripple laminations. Rip-up clasts occur locally along the scoured base. Locally, millimeter scale root traces and plant imprints that are 5-20 cm are present in the upper 0.2 m of the sandstones. As in Sub-unit 2a, lateral accretion surfaces are not identified within Sub-unit 2b. The siltstone and mudstone deposits separating the sandstones are poorly laminated to non-laminated and contain abundant plant fragments and fern imprints. The siltstones and mudstones are dark brown and appear featureless.

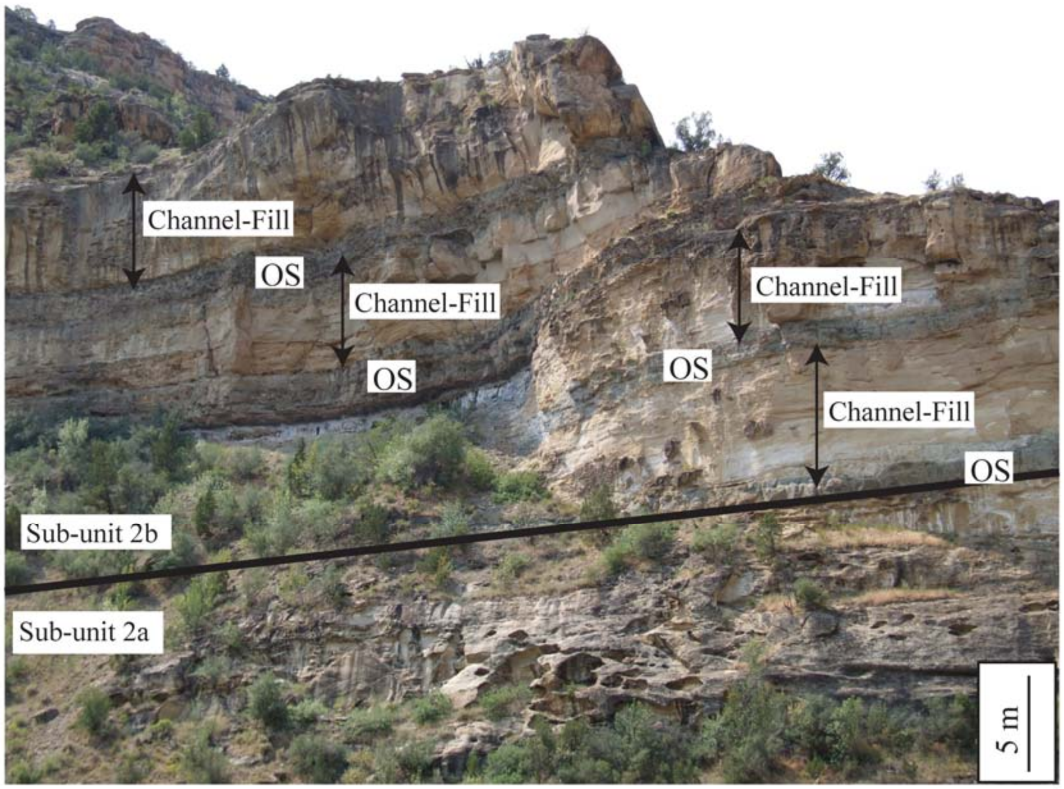


Figure 12: Stacked, laterally continuous, single-story channel-fill sandstones separated by overbank siltstones (OS) in Sub-unit 2b. Each channel-fill sandstone becomes finer grained from the base to the top of the sub-unit.

Sub-unit 2b represents a vertical succession of laterally amalgamated single-story, channel-fill sandstones that are separated by overbank deposits. The grain size trend seen in Sub-unit 2b indicates that the energy of the fluvial systems decreased throughout deposition of Sub-unit 2b, resulting in an overall decrease in average grain size in each channel-fill sandstone. The compound and complex cross stratification indicates that the fluvial system was dominated by downstream bar form migration typical of braided stream deposits (Miall, 1996). The presence of multiple, laterally amalgamated channel-fill deposits also suggests deposition by a fluvial systems with multiple channels such as a braided stream deposit (Miall, 1996; Bridge, 2006). The

Figure 13: Fining upward channel-fill succession of Sub-unit 2b. Left side of photo shows soft sediment deformation. Right side of figure shows tabular and trough cross-stratification. The thickness of the cross stratification decreases vertically from the base of the channel-fill to the top.

absence of lateral accretion surfaces suggests that the fluvial system was not highly meandering. Average grain size differences between the channel-fill deposits in Sub-unit 2a and Sub-unit 2b indicate that the fluvial systems in Sub-unit 2b are lower energy than those in Sub-unit 2a. The channel-fill sandstones represent fluvial systems with multiple channels such as a braid plain (Miall, 1996; Bridge 2006). Root traces and abundant plant imprints indicate that the floodplain was vegetated.

Sub-unit 2c

Sub-unit 2c directly overlies Sub-unit 2b and consists of aggradationally stacked, isolated, lens shaped sandstones encased in siltstones and mudstones (Fig. 14). The total thickness of the sub-unit ranges from 4 to 16 m and it extends laterally across the length of the exposure (up to six kilometers). The sandstones are 0.3 to 0.8 m in thickness and thin laterally to less than 0.1 m thick. Where observed, individual lens shaped sandstones extend 3 to 20 m laterally, although poor exposures and weathering makes lateral correlation difficult. The grain size of the sandstone lenses ranges from fine lower to very fine upper with no vertical grain size trend identified. Sandstone lenses have a scoured base and are current-ripple dominated or appear structureless, likely due to weathering. Locally, sandstones appear bioturbated and contain millimeter scale root casts and fern imprints in the upper five centimeters (Fig. 15). The siltstones and mudstones that encase the sandstones are dark brown and appear massive. They contain abundant plant fragments and fern imprints (Fig. 15).

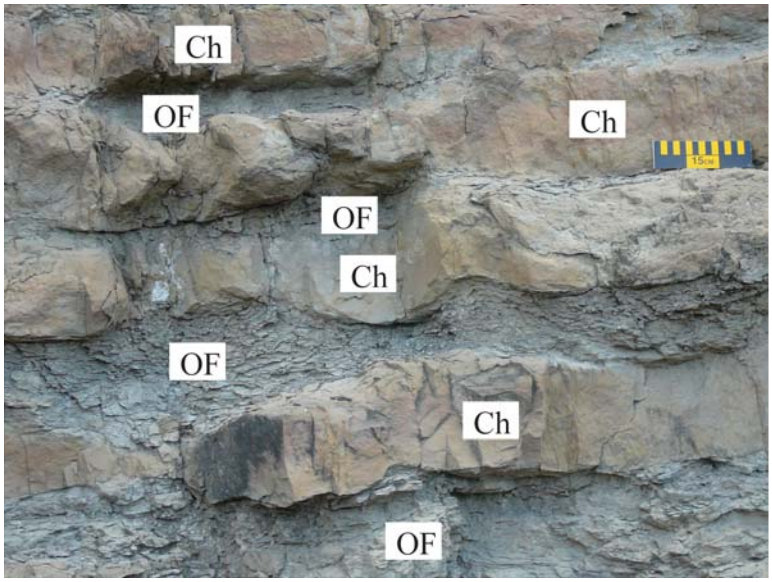


Figure 14: Small, isolated channel-fill sandstones (Ch) encased in overbank fines (OF) in Sub-unit 2c.

Sub-unit 2c is interpreted to represent deposition from small, low energy streams that flowed through a vegetated floodplain. The dark brown color of the overbank deposits suggests that the floodplain was poorly drained (Potter et al., 2005). Sub-unit 2c contains the lowest energy channel-fills within Unit 2.

Sequence Stratigraphy of Unit 2

The vertical relationship of the three sub-units within Unit 2 is distinctive and repetitive and defines a stratal package (Fig. 9; Fig. 16; Fig. 17). Stratal packages are laterally continuous for the extent of the exposure (up to 10 km). Sixteen stratal packages occur within Unit 2. Twelve of the stratal packages are complete and four are incomplete with Sub-unit 2c or Sub-units 2b and 2c absent. In these instances, Sub-unit 2a or Sub-unit 2b is truncated by an erosional surface, which in turn is overlain by Sub-unit 2a. In these instances the sequence boundary is identified by an abrupt increase in the grain size from the scour base of the lower Sub-unit 2a to the scour base of the upper Sub-unit 2a.

The boundary between each of the complete stratal packages is sharp with an abrupt change from Sub-unit 2c of the underlying package to Sub-unit 2a of the overlying package. An abrupt increase in the grain size of channel-fill deposits marks

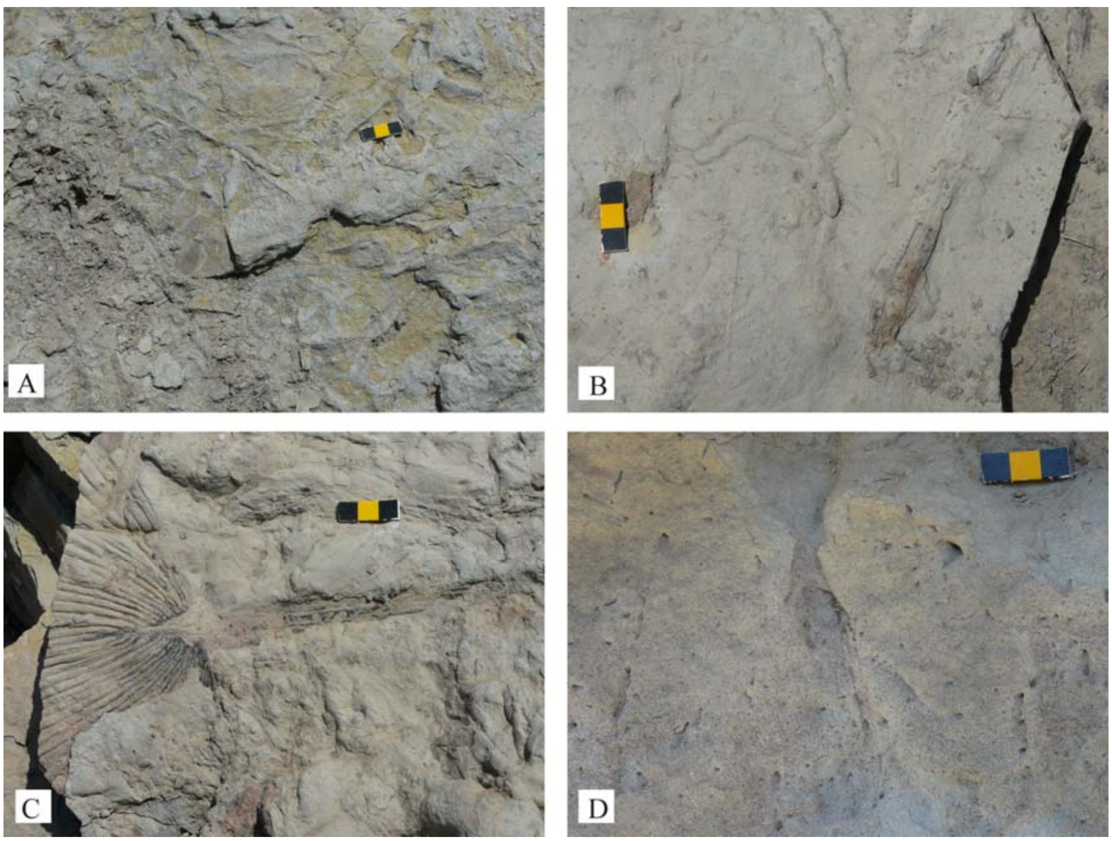


Figure 15: Several features from Sub-unit 2c. A and B) Trace fossils present on the upper surface of a siltstone bed. C) Plant imprint found on the same surface as the trace fossils. D) Millimeter scale root traces present in the upper 10 cm of a very-fine grained channel-fill deposit. Each photo has a 3cm scale.

the boundary, along with an increase in cross bed thickness and a facies change from abundant overbank deposits preserved below the boundary to channel-fill sandstones directly above the boundary. These changes indicate an abrupt increase in energy and decrease in accommodation. The surface across which these changes occur is interpreted to be a sequence boundary (Fig. 9). Each stratal package is bounded by sequence boundaries.

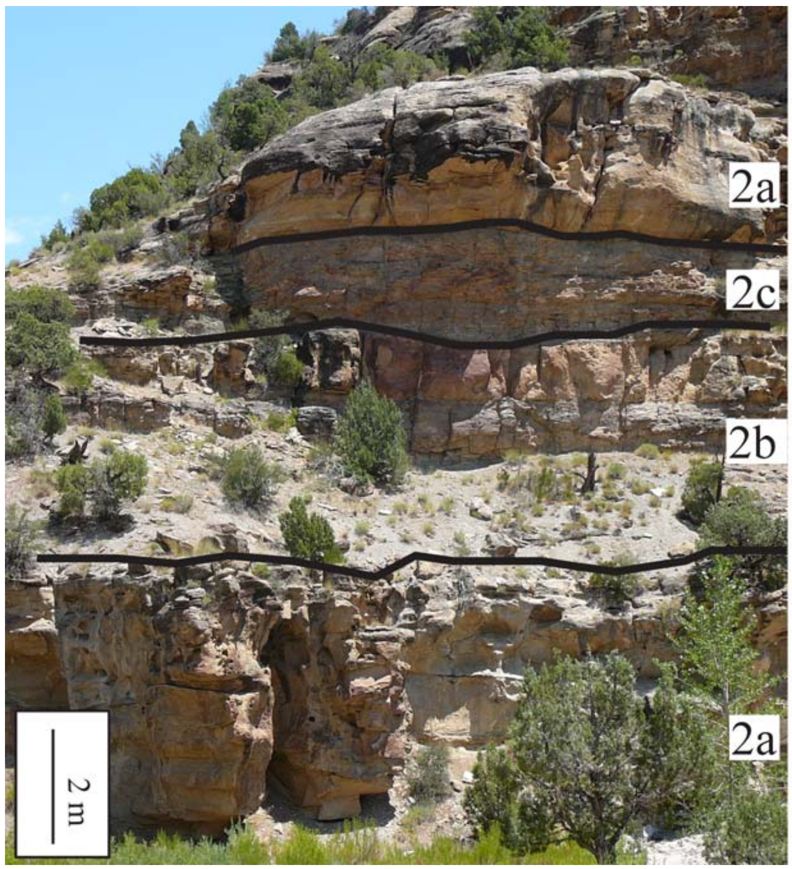


Figure 16: One of the thinner complete stratal packages in Unit 2 showing Sub-units 2a, 2b and 2c and Sub-unit 2a of the overlying stratal package. Note vertical decrease in sandstone thickness and increase in overbank preservation. Sub-unit boundaries are in black.

Figure 17: Unit 2 showing Sub-units 2a, 2b, 2c and Sub-unit 2a from an overlying sequence. Red line represents a sequence boundary. Black lines show boundaries between sub-units. The yellow lines show the thickness of sandstones at their thickest location within the photo. Photo shows how sandstones in Sub-unit 2b pinch and swell. The thickness of sandstones within Sub-unit 2b decreases vertically when looking at the thickest portions of each sandstone.

Each stratal package displays a vertical trend characterized by an overall upward decrease in gradient of the fluvial system and an increase in accommodation from Sub-unit 2a to Sub-unit 2c. In each stratal package there is a vertical decrease in average grain size of the channel-fill sandstones, a decrease in cross bed thickness and a decrease in channel-fill sandstone thickness. These vertical trends indicate a decrease in fluvial energy as a result of a decrease in the gradient of the fluvial system (Richards, 2006). Throughout the stratal package there is a vertical increase in overbank preservation and a decrease in channel-fill connectivity representing a change from low accommodation at the base of the stratal package to higher accommodation at the top (Shanley and McCabe, 1994).

Each stratal package is interpreted to be a depositional sequence, which initiates with a basal sequence boundary overlain by a lowstand systems tract (Sub-unit 2a), a transgressive systems tract (Sub-unit 2b) and a highstand systems tract (Sub-unit 2c) (Fig. 9). The early lowstand is represented by the basal sequence boundary where sediment bypassed the system and incision occurred in response to a fall in base level and an increased fluvial gradient. Deposition of Sub-unit 2a occurred during the late lowstand as base level slowly began to rise. Within each depositional sequence, Sub-unit 2a contains the highest energy fluvial deposits as indicated by the largest grain size and cross bed thickness. The absence of preserved overbank deposits indicates that the fluvial system had time to rework lateral deposits. The aggradational and nested nature of the stacked channel-fill deposits indicates that base-level rise was slow, steady and continuous. Slow and continuous base-level rise results in the slow and continuous formation of accommodation and is indicative of late lowstand conditions (Richards, 1996). The nested channel-fill complexes with interconnected channel-fill sandstones are characteristic of late lowstand deposits. Similar deposits described in the Cretaceous Castlegate Sandstone of Utah (Van Wagoner et al., 1995; Yoshida et al., 1996) and in the Cretaceous Mesa Rica Sandstone of New Mexico (Holbrook and Dunbar, 1992) have been interpreted as lowstand deposits. Both cases describe sheet sandstones consisting of vertically and laterally amalgamated channel-fill deposits that are interpreted as braided river deposits in a lowstand systems tract.

Sub-unit 2b is characterized by a vertical decrease in fluvial energy indicated by the overall decrease in average grain size and cross bed thickness from the lowest channel-fill sandstone to the highest. This occurs in response to an increase in the rate of base-level rise associated with the transgressive systems tract. As base-level rises, the fluvial gradient decreases resulting in lower energy fluvial systems. The change to interbedded, single-story channel-fill sandstones and overbank fines in Sub-unit 2b is interpreted to represent an increase in accommodation due to an increase in the rate of base-level rise, which is characteristic of the shift from the lowstand systems tract to the transgressive systems tract (Richards, 1996). Each laterally continuous channel-fill sandstone within Sub-unit 2b is interpreted to be a parasequence with the top of each single-story channel-fill representing a parasequence boundary. Each vertically stacked parasequence within the sub-unit is progressively thinner and contains a smaller average grain size than the underlying parasequence. This pattern represents a backstepping parasequence stacking pattern common in transgressive systems tract deposits. The backstepping stacking pattern indicates that the rate of base-level rise outpaced the rate of sediment supply. The presence of well-developed parasequences (i.e. not nested and amalgamated channel-fills such as those in Sub-unit 2a) indicate that base-level rise was punctuated with still stands allowing time for the reworking of lateral deposits by the fluvial system. Overbank deposits are preserved between vertically stacked channel-fill sandstones and represent times when base-level rise was more rapid. Similar deposits in the Turonian to Campanian Straight Cliffs Formation of the Kaiparowits Plateau (Shanley and McCabe, 1995) and the Campanian Castlegate Sandstone in Utah (Olsen et al., 1995) are interpreted as transgressive systems tract deposits.

The lowest energy fluvial deposits within the sequence are in Sub-unit 2c and represent deposition during the highstand systems tract. Fluvial systems with the lowest gradient in the depositional sequence occur during deposition of Sub-unit 2c. The top of each channel-fill sandstone, and its laterally equivalent overbank deposits, represents a parasequence boundary. The channel-fill deposits in Sub-unit 2c are the thinnest channel-fill deposits within the depositional sequence. These deposits are indicative of a system with decreasing accommodation resulting in thinner parasequences. Although parasequences are thinner, parasequence development occurs quickly resulting in isolated channel-fill sandstones with limited lateral continuity. Vertically isolated channel-fill deposits encased within overbank fines, described and interpreted as highstand deposits, occur in the Turonian to Campanian Straight Cliffs Formation of the Kaiparowits Plateau of Utah (Shanley and McCabe, 1995). The highstand systems tract deposits of the Straight Cliffs Formation, like those in the study area, overlie transgressive systems tract deposits and are truncated by a sequence boundary indicating a subsequent drop in base level. The vertical stacking of sixteen of these depositional sequences reflects repetitive rises and falls in base level during deposition of Unit 2.

Unit 3

Unit 3, as with Unit 2, is divided into three sub-units: Sub-unit 3a consists of laterally continuous stacked sandstones; Sub-unit 3b consists of isolated lens shaped sandstones encased in mudstones and siltstones; and Sub-unit 3c consists of mottled sandstones to mudstones. Unit 3 is laterally continuous for the length of the outcrop (up to three kilometers).

Sub-unit 3a

Sub-unit 3a ranges in thickness from 8 to 22 m and consists of laterally continuous sandstones interbedded with siltstones and mudstones. The sandstones are generally separated by up to two meters of mudstone and siltstone but in some localities the sandstones are vertically amalgamated. Each sandstone has a scoured base and fining upward succession. Within Sub-unit 3a, one to three of these sandstones are vertically stacked. Within the sub-unit there is a vertical decrease in sandstone thickness with the lowest sandstone ranging from six to eight meters in thickness and the highest sandstone ranging from four to six meters in thickness. As in Sub-unit 2b, two-grain size trends are present within Sub-unit 3a. Each individual sandstone displays an upward-fining profile from medium-upper to medium-lower at the base to fine lower to very-fine upper at the top. The second grain size trend is an overall decrease in average grain size from the lowest sandstone in the sub-unit to the highest. The average grain size of the sandstones in Sub-unit 3a is the largest average grain size within the entire formation. Similar to Sub-unit 2b, Sub-unit 3a has compound and complex bedding resulting in large scale cross beds (0.2 to 1.2 m thick) overlain by current ripples. There are two trends in cross bed thickness. First, within each sandstone the cross bed thickness decreases vertically from the scoured base to the top (from ~0.2 to ~1.2 m at the base to~0.1 to ~0.2 m at the top).

Secondly, there is an overall decrease in average cross bed thickness vertically from the lowest sandstone in the sub-unit to the highest sandstone (the larger cross beds at the base of the sandstone in the lower sandstones is ~1 to ~1.5 m while the higher sandstones have cross beds only as large as ~0.1 to ~0.4 m). Current ripples dominate the uppermost 10 to 20 cm of each sandstone. Lateral accretion surfaces are absent. Chert and quartz pebbles occur locally at the scoured base and within troughs. Convolute bedding occurs locally. In places, the top three to five centimeters of the sandstones contain millimeter scale root traces. The mudstones and siltstones of Sub-unit 3a are brown to reddish brown and lighter in color than those of Unit 1 or 2. The siltstones and mudstones appear structureless and contain disseminated plant fragments.

Sub-unit 3a represents stacked, channel fill sandstones separated by overbank deposits. The coarse grain size of the deposits sometimes reaching pebble size clasts, indicates the highest energy fluvial system within the Williams Fork Formation. Fluvial channels are dominated by downstream barform migration. The depositional environment for Sub-unit 3a is similar to that of Sub-unit 2b representing a braid plain (Miall, 1996). The presence of rooting in overbank deposits suggests that there was time for vegetation to form lateral to the fluvial deposits. The reddish color of the overbank deposits indicates that the floodplain during deposition of Sub-unit 3a was better drained than the floodplain during deposition of Units 1 or 2 (Potter et al., 2005).

Sub-unit 3b

Sub-unit 3b directly overlies Sub-unit 3a and consists of aggradationally stacked isolated sandstone lenses encased in mudstone and siltstone, similar to Sub-unit 2c. Overall the sub-unit ranges in thickness from three to six meters and extends laterally for the length of the exposure (up to one kilometer). The sandstone lenses are 0.3 to 0.4 m in thickness and thin laterally to less than 0.1 m in thickness. The sandstone lenses extend laterally 10 m or less. Grain size of the sandstone lenses is very fine lower to fine lower and no grain size trend is identified. Each sandstone lens has a scoured base and locally the sandstones contain current ripple laminations, but extensive weathering of the sub-unit results in many of the sandstones appearing structureless. Several of the sandstone lenses contain millimeter scale root traces within the top 10 cm. The mudstones and siltstones that encase the sandstones are brownish-red to green and locally have a mottled texture. The mudstones and siltstones appear structureless and locally contain plant fragments.

Sub-unit 3b represents low energy channel fill deposits and associated overbank deposits similar to those of Sub-unit 2c. The color of the mudstone is typical of a well-drained floodplain while the mottled texture suggests bioturbation and the early stages of soil formation (Potter et al., 2005). These deposits are likely analogous to the deposits of low energy streams flowing through a well-drained, vegetated floodplain. The presence of rooting at the top of the channel-fill deposits indicates that abandoned channels became vegetated.

Sub-unit 3c

Sub-unit 3c directly overlies Sub-unit 3b, and consists of one or two stacked, upward-fining successions from very-fine lower sandstone to mudstone. The sub-unit ranges from one to four meters, and extends laterally the length of the exposure (up to 0.5 km). Each upward-fining profile displays a change in color from red-brown at the base grading up into purple at the top (Fig. 17). Mottling is prevalent throughout with a green-gray color to the mottles. There is also an increase in the amount of weathering from the base to the top. Sub-unit 3c contains abundant ped structures. Near the base of the upward-fining succession these ped structures are blocky while near the top the ped structures become granular to crumby (Soil Survey Staff, 1975; Retallack, 2001). Millimeter scale root traces are prevalent throughout the sub-unit but are most abundant at the top of the upward-fining profile (Fig. 17).

Sub-unit 3c represents soil development with each upward-fining profile representing a paleosol. One of the best criteria for the interpretation of paleosols are the presences of root traces (Retallack, 2001), which are prevelant throughout Sub-unit 3c. The red-brown to purple color of the mudstone is indicative of continental rocks and represents the oxidation of iron oxides in a well-drained floodplain (Potter et al., 2005). The green mottling is an indicator of bioturbation and the alternating wetting and drying of sediment during soil formation (Retallack, 2001; Potter et al., 2005). The fining upward profile may be a result of increased weathering from the base of the paleosol to the top or it may reflect a characteristic of the parent rock, such as the upward-fining profile of a channel-fill sandstone. There are, however, no sedimentary structures that indicate that the parent rock was a channel-fill deposit.

Sequence Stratigraphy of Unit 3

Unit 3, similar to Unit 2, has a predictable and repetitive arrangement of sub-units, defining a stratal package (Fig. 9; Fig. 18). Three of these stratal packages are recognized in Unit 3. The stratal packages range in thickness from 16 to 30 m and, on average, are thinner than the stratal packages in Unit 2.

The boundary between each stratal package is sharp with an abrupt change from the paleosols of Sub-unit 3c to the high-energy fluvial systems of Sub-unit 3a. This boundary is interpreted as a sequence boundary. As in Unit 2, accommodation is interpreted to increase and the gradient of each fluvial system is interpreted to decrease vertically from the base to the top of each stratal package. These trends are indicated by the vertical decrease in fluvial energy and the vertical increase in overbank preservation. As in Unit 2, each stratal package within Unit 3 represents a depositional sequence with a basal sequence boundary, a lowstand to transgressive systems tract (Sub-unit 3a) and a highstand systems tract (Sub-units 3b and 3c) (Fig. 6).

The basal sequence boundary marks the beginning of the lowstand as base-level falls and fluvial gradient increases, resulting in sediment bypass and fluvial incision. In Unit 3, the lowstand is represented only by the sequence boundary and a well-developed late lowstand systems tract deposit, such as those seen in Unit 2, is

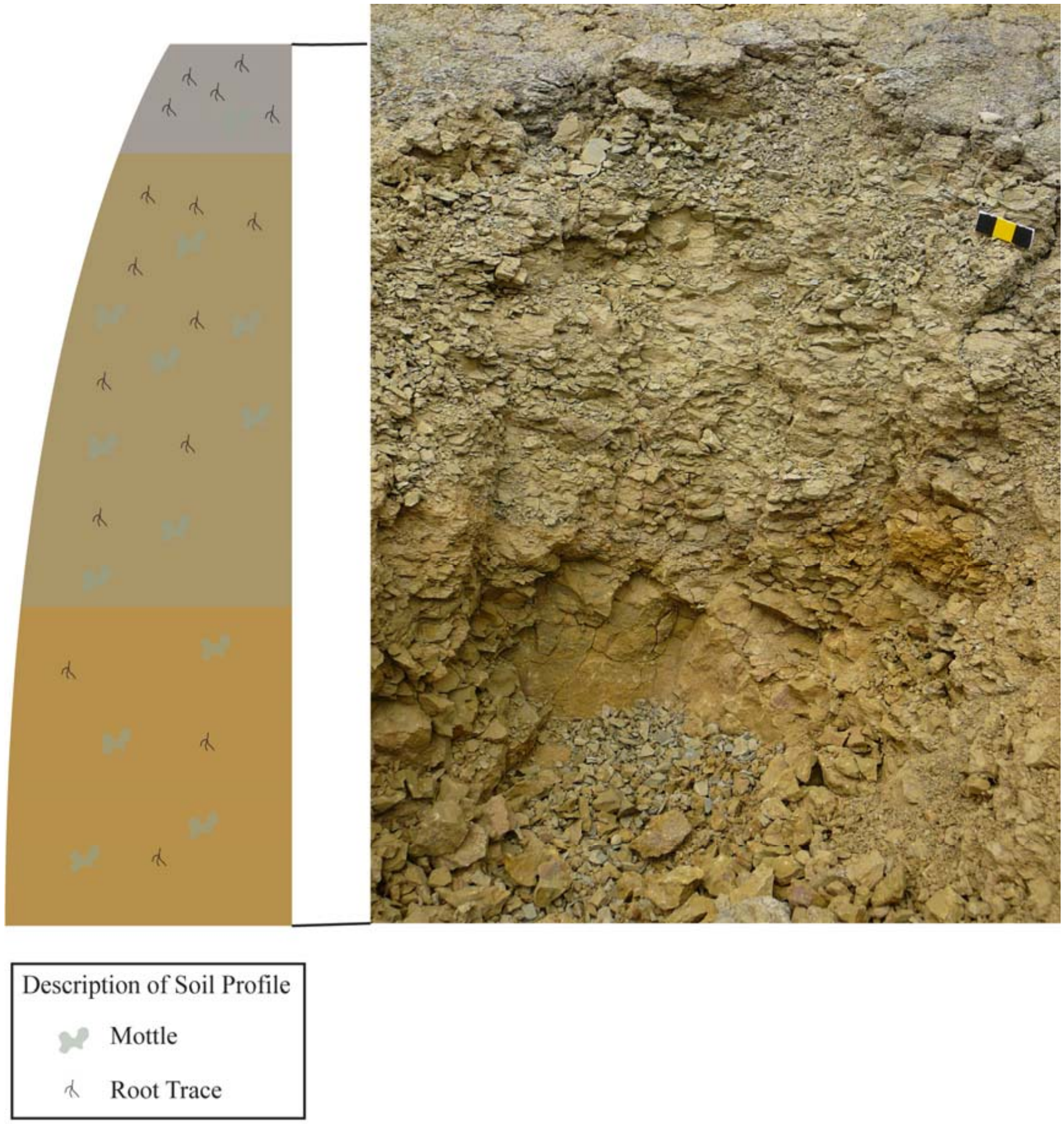


Figure 17: The photograph shows a paleosol horizon from Sub-unit 3c. The section on the left is a weathering profile displaying features of the soil horizon that do not show up well in the photograph. There is a color change from the base to the top of the profile. The amount of root traces increases vertically from the base to the top of the profile. Motteling is present throughout. The photo shows the presence of blocky ped structures near the base and granular to crumby ped structures near the top. The scale in the photograph is 3 cm.

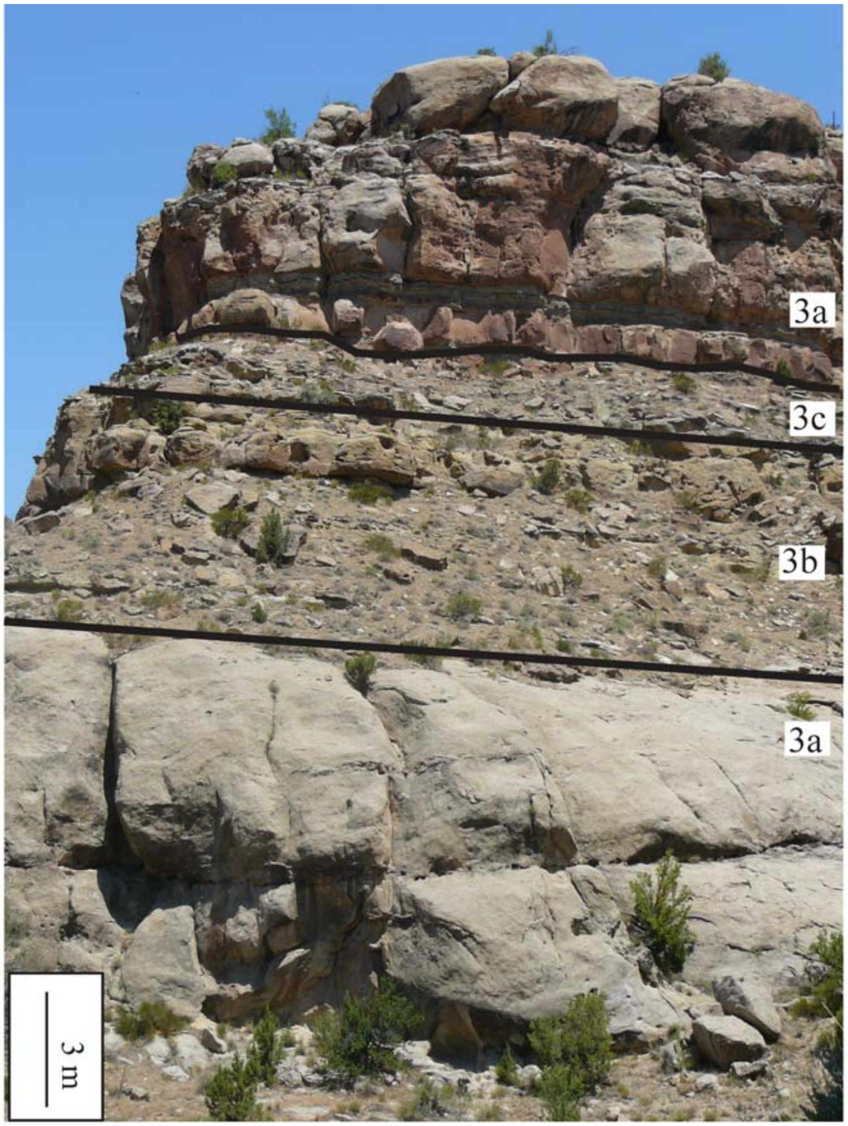


Figure 18: Stratal package from Unit 3 showing Sub-units 3a, 3b and 3c. The base of the stratal package is not shown and a lower sandstone from Sub-unit 3a is not pictured. Black lines separate sub-units. Overall increase in overbank preservation and decrease in sandstone thickness vertically through the stratal package is shown.

absent. Instead, the single-story channel-fill sandstones of Sub-unit 3a sit directly on the sequence boundary. The top of each channel-fill represents a parasequence boundary and the parasequences display a backstepping parasequence stacking pattern indicated by the decrease in fluvial energy from the base to the top of the sub-unit. The decrease in parasequence thickness from the base to the top of the sub-unit represents a decrease in accommodation associated with the decrease in the rate of base-level rise. The well-developed nature of the parasequences once again indicates that base-level rise was punctuated by stillstands. As with the deposits of Sub-unit 2b, the single-story channel-fill sandstones separated by overbank deposits of Sub-unit 3a represent deposition during the transgressive systems tract. Unlike Sub-unit 2b, however, the channel-fill sandstones of Sub-unit 3a are amalgamated in several localities indicating accommodation was lower for Sub-unit 3a than Sub-unit 2b.

The lowest energy fluvial systems within each sequence occur in Sub-unit 3b, and they represent the lowest gradient fluvial systems within the depositional sequence. As in Sub-unit 2c, Sub-unit 3b was deposited during a highstand systems tract with a higher accommodation-to-sediment supply ratio than in the underlying sub-units. This resulted in the deposition of isolated, aggradationally stacked, channel-fill deposits typical of highstand systems tracts. Paleosols of Sub-unit 3c occur at the top of each depositional sequence and are typical of the late highstand where soil maturity is often at its highest (Wright and Marriott, 1993) The top of each paleosol represents a parasequence boundary (Kamola and Van Wagoner, 1994). These paleosols indicate that accommodation was low as the rate of base-level rise slowed allowing enough time to weather deposits and form soils. There are at least two explanations for why well-developed soils occur in the highstand deposits of Unit 3 but not in the highstand deposits of Units 1 or 2. It is possible that accommodation rates were much higher in Units 1 and 2, resulting in poorly drained floodplains that did not have the right conditions for soil development. This theory is consistent with the dark color of the overbank deposits in Units 1 and 2 that typically indicate poorly drained floodplains (Potter et al., 2005). A second hypothesis is that soil development did occur in Unit 2 but subsequent erosion represented by the overlying sequence boundary removed the late highstand deposits.

CONCLUSIONS

Stratal patterns within the Williams Fork Formation record changes in accommodation throughout the final stage of fill of the Cretaceous Western Interior foreland basin. These stratal patterns are analyzed and interpreted using a sequence-stratigraphic approach. The sequence-stratigraphic interpretation of each of the three units is possible because fluvial systems respond to changes in base level in predictable ways. These changes in base level are identified by analyzing the lateral and vertical changes in fluvial facies throughout the thick fluvial deposits.

The Williams Fork Formation is divided into three units based on the changes in stratal packaging. Unit 1 contains no distinct stratal pattern while Units 2 and 3 both contain repetitive stratal packages interpreted as depositional sequences. Unit 1 consists of isolated channel-fill sandstones encased in overbank fines. The top of each channel-fill represents the top of a parasequence and the multiple parasequences are aggradationally stacked within the unit. Unit 1 represents deposition during one or multiple highstand systems tracts.

Unit 2 contains sixteen repetitive stratal packages. Each stratal package is bounded by surfaces of regional erosion and displays a vertical change in fluvial style that is interpreted to reflect an overall decrease in fluvial energy and an increase in accommodation from the base of the stratal package to the top. Each stratal package represents a depositional sequence with a basal sequence boundary overlain by a lowstand, transgressive and highstand systems tract. In several cases the highstand systems tract or the transgressive and highstand systems tracts are absent indicating that the depositional sequence is not fully preserved.

Unit 3 contains three repetitive stratal packages. As in Unit 2, each stratal package is bounded by surfaces of regional erosion and displays a vertical change in fluvial style that is interpreted to reflect an overall decrease in fluvial energy and an increase in accommodation from the base of the stratal package to the top. Each stratal package represents a depositional sequence. The depositional sequences in Unit 3 are different than those in Unit 2 because they do not contain late lowstand systems tract deposits. Instead each depositional sequence in Unit 3 consists of a basal sequence boundary, a lowstand to transgressive systems tract and a highstand systems tract.

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CHAPTER THREE

Eustatic and Tectonic Control of Stratal Packaging in the Upper Cretaceous Williams Fork Formation, Piceance Basin, Colorado

ABSTRACT

The Cretaceous Williams Fork Formation consists of fluvial and related deposits that represent the final stage of fill within the Cretaceous Western Interior foreland basin. The formation contains distinct stratal packages that represent changes in accommodation through time. The formation is divided into three units based on the facies and stratal patterns. The stratal patterns in each unit are analyzed using sequence stratigraphy. Unit 1 consists of aggradationally stacked channel-fill sandstones isolated within overbank deposits and represents deposition during one or multiple highstands. Units 2 and 3 contain stacked, repetitive stratal packages interpreted as depositional sequences. Each sequence is bracketed by basal a sequence boundary and contains a lowstand, transgressive and highstand systems tract.

The changes in accommodation reflected in the depositional sequences are proposed to be controlled by subsidence, formed from tectonic and sediment loading, and eustatic changes within the foreland basin. Tectonic loading resulting in subsidence is proposed to control the formation of accommodation allowing for the stacking of multiple fluvial successions. Eustatic changes are proposed to control higher order changes in accommodation reflected in individual stacked depositional sequences present in Units 2 and 3. When the rate of eustatic fall outpaced the rate of tectonic subsidence, a sequence boundary was formed.

INTRODUCTION

The Cretaceous Williams Fork Formation, deposited in the Cretaceous Western Interior foreland basin, displays distinct stratal packages (Chapter 2). These stratal packages record the fill of the foreland basin through time are not random but display repetitive patterns. The presence of these patterns reflects a single or multiple driving force(s) that control accommodation during deposition of the formation.

The Williams Fork Formation is bounded below by the marine Iles Formation and is overlain by the fluvial and lacustrine Wasatch Formation. Ages for deposition of the Williams Fork Formation are uncertain: interpretation of palynologic data places the base of the formation in the latest Campanian to the early Maastrichtian and the top of the formation deposition at the latest Maastrichtian (Johnson and May, 1980). Interpretations by Johnson and May (1980) indicate that deposition of the formation did not continue after the end of the Cretaceous. There are no other dating techniques that have been applied to the formation to help narrow down the time constraints. If the longest amount of time interpreted from the palynological date is ~14m.y. from ~70 Ma to ~66 Ma. The shortest amount of time for deposition of the formation interpreted from the palynological data is ~8 m.y. from ~74 Ma to ~66 Ma. This study focuses on outcrops approximately twenty miles east of Grand Junction, Colorado (Fig. 1).

This chapter begins by describing the stratal packages and interpreting them using a sequence-stratigraphic approach. Next is a discussion of the mechanism(s) responsible for forming overall accommodation within the basin that allowed the

Williams Fork Formation to be preserved. The sequence stratigraphic interpretation shows that accommodation was not formed continuously within the basin. There were times of deposition and times of erosion. The accommodation history of each unit is discussed focusing on what mechanism(s) are responsible for shorter-term changes in accommodation. The interaction between the shorter-term and longer-term changes in accommodation result in the stratal packages present. Finally, the overall changes in accommodation throughout the entire formation are discussed.

STRATAL PACKAGES

Previous work identified and interpreted stratal packaging within the Williams Fork Formation. Changes in accommodation were defined by interpreting stratal packages using a sequence-stratigraphic approach. The following section summarizes the descriptions and sequence-stratigraphic interpretations of the stratal packaging from Chapter Two.

Two scales (large and small) of stratal packaging are identified within the Williams Fork Formation. The larger scale stratal packaging divides the formation into three units based on facies and smaller scale stacking patterns: Unit 1 (basal ~130 meters), Unit 2 (middle ~320 meters), and Unit 3 (upper ~70 meters) (Fig. 2). At this scale, an overall change in accommodation style is seen between the three units. The smaller scale stratal packages are repetitive, stacked stratal patterns present within Units 2 and 3 (Fig. 2). These stratal patterns record repetitive, cyclic changes in base level with time. Accommodation is defined as the space available for sediment to

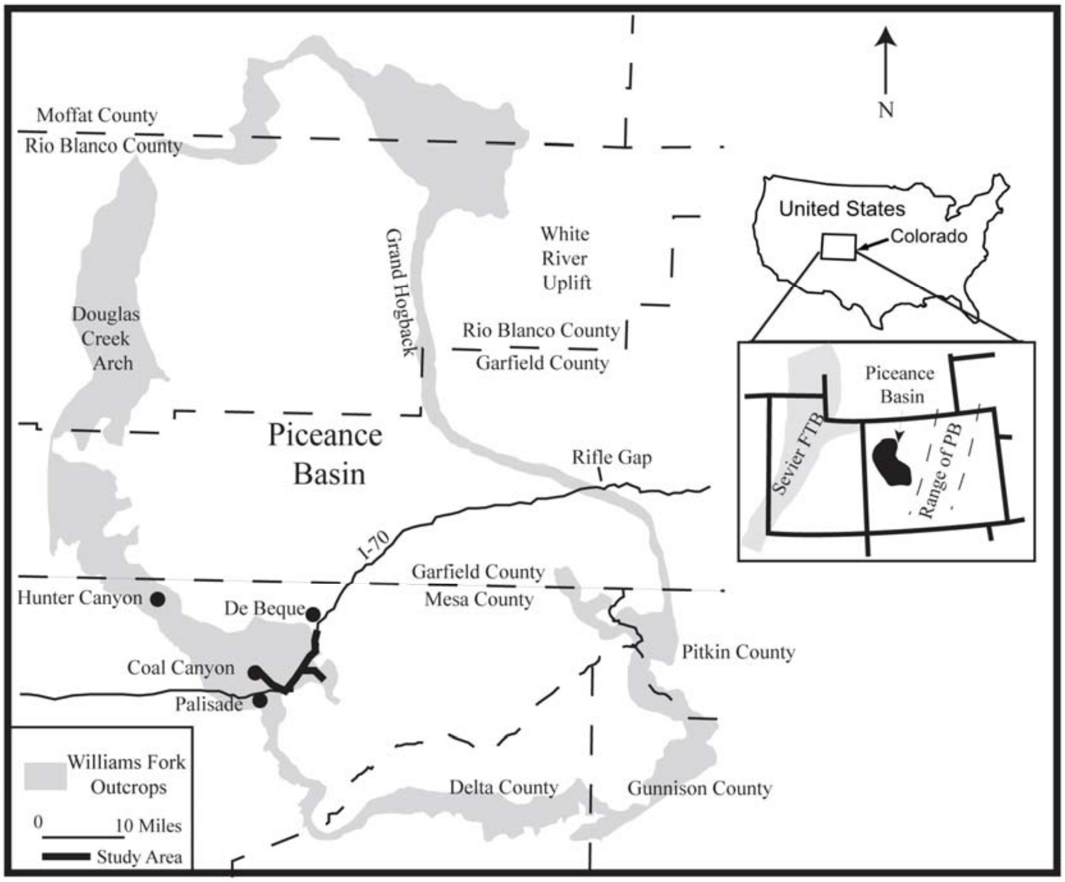
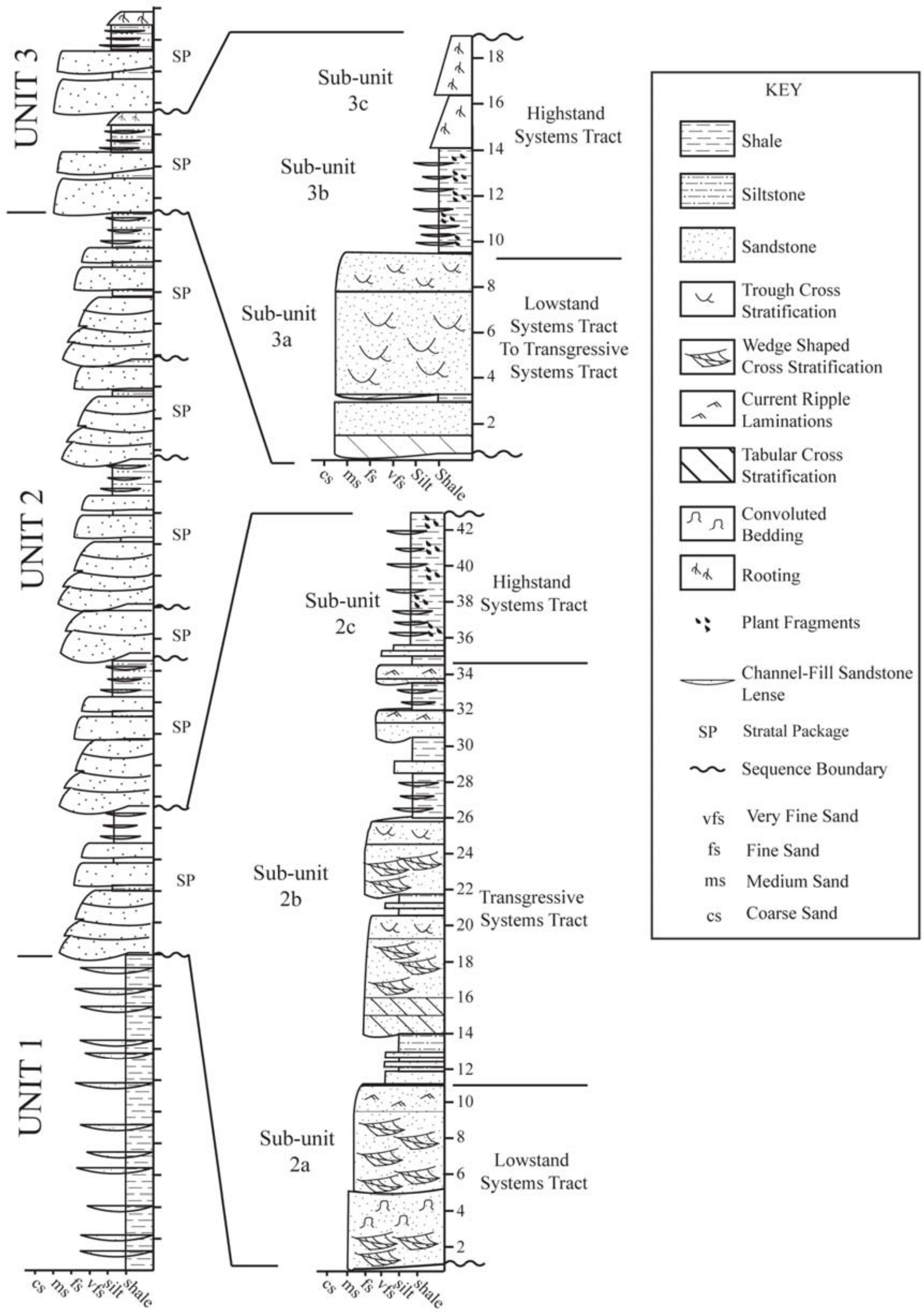


Figure 1: Map showing location of the Williams Fork Formation outcrops within the Piceance Basin (Colorado) and location of this study (Modified from Pranter et al., 2007).

accumulate (Jervey, 1988). Accommodation can be interpreted from stratal patterns seen in fluvial strata and is controlled by base level. In fluvial systems, low connectivity of channel-fill sandstones and abundant overbank fines indicate high accommodation while high connectivity of channel-fill sandstones and scarce overbank fines indicate low accommodation (Shanley and McCabe, 1995).

Figure 2: (On next page) Generalized stratigraphic section of Williams Fork Formation showing Units 1 through 3 and detailed stratigraphic sections of representative stratal packages from Units 2 and 3 with sequence stratigraphic interpretation. Schematic does not show actual thickness or number of sequences present within each Unit. Vertical axis shows thickness in meters.



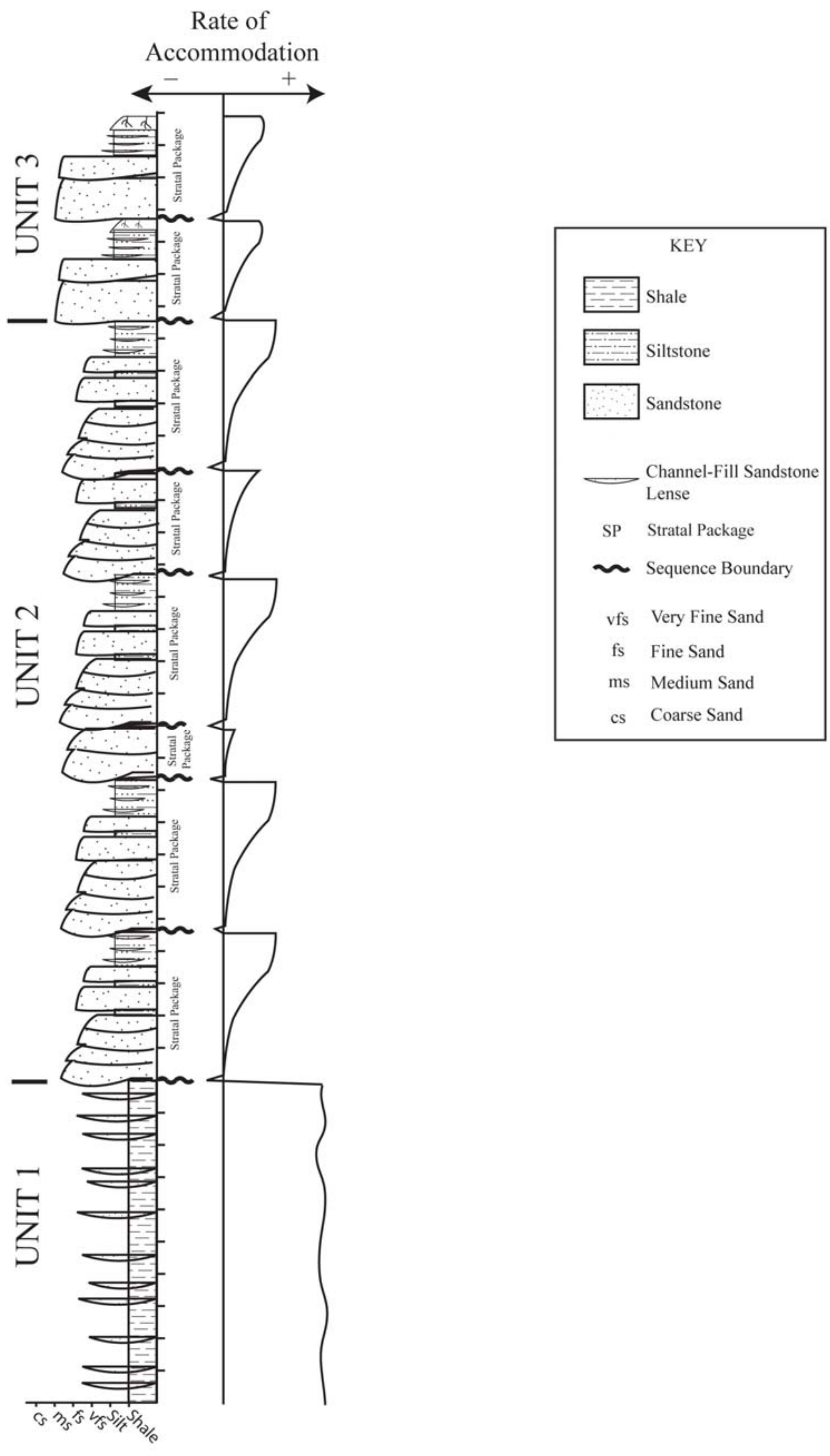
UNIT 3
UNIT 2
UNIT 1
SP
Sub-unit 3c
Sub-unit 3b
Sub-unit 3a
Highstand Systems Tract
Lowstand Systems Tract To Transgressive Systems Tract
Sub-unit 2c
Highstand Systems Tract
Sub-unit 2b
Transgressive Systems Tract
Sub-unit 2a
Lowstand Systems Tract
cs ms fs vfs silt shale
cs ms fs vfs Silt Shale
KEY
Shale
Siltstone
Sandstone
Trough Cross Stratification
Wedge Shaped Cross Stratification
Current Ripple Laminations
Tabular Cross Stratification
Convoluted Bedding
Rooting
Plant Fragments
Channel-Fill Sandstone Lense
SP Stratal Package
Sequence Boundary
vfs Very Fine Sand
fs Fine Sand
ms Medium Sand
cs Coarse Sand

Unit 1 consists of isolated, single-story channel-fill sandstones encased in overbank fines and represents a period of high accommodation. Channel-fill sandstones are lens shaped with a scoured base and a lateral extent of no more than ~ 800 m. The abundance of overbank fines is interpreted to represent abundant flooding events associated with low-gradient fluvial systems with high suspended load to bed load ratios (Leopold and Wolman, 1957). The top of each channel-fill sandstone, and its associated overbank deposits, represents the top of a parasequence. Each parasequence represents a period of new accommodation in the basin. Parasequences are aggradationally stacked within Unit 1. Channel-fill attributes such as grain size and cross-bed dimensions remain constant from parasequence to parasequence. Smaller scale stratal patterns, like those present in Units 2 and 3, are not identified within Unit 1. Unit 1 is interpreted to be deposited within the highstand systems tract. Sequence boundaries are not identified within Unit 1, but may be present as interfluv expressions of sequence boundaries. Criteria used to identify sequence boundaries include (1) an abrupt increase in channel-fill grain size and cross-bed set thickness associated with an increase in the gradient and energy of the fluvial system and (2) an abrupt increase in channel-fill sandstone connectivity reflecting decreased accommodation.

Unit 2 overlies Unit 1 and contains distinct, repetitive stratal patterns (stratal packages are not random). When a complete stratal package is present it is divided into three sub-units: Sub-unit 2a (basal unit; laterally continuous nested channel-fill sandstone complex), Sub-unit 2b (single-story channel-fill sandstones separated by overbank fines), and Sub-unit 2c (succession of small scale, aggradationally stacked, isolated channel-fill sandstones that are encased in overbank fines) (Fig. 2). Fluvial gradient in each sub-unit decreases up section, and each stratal package records an increase in accommodation from Sub-unit 2a to Sub-unit 2c (Fig. 3). Each stratal package is interpreted to represent a depositional sequence bounded by sequence boundaries and contains a lowstand systems tract (Sub-unit 2a), a transgressive systems tract (Sub-unit 2b), and a highstand systems tract (Sub-unit 2c) (Fig. 2). The sequence boundaries within Unit 2 are interpreted to be fourth-order sequence boundaries and each depositional sequence is estimated to represent ~0.4 to 0.6 Ma. The timing of the fourth order sequences was estimated by dividing the total amount of time allowed for deposition (8-13 Ma) by the number of identified depositional sequences and then subtracting the amount of time represented by Unit 1. The amount of time represented by Unit 1 was estimated using average sedimentation rates for the distal portions of foreland basins. These estimations are based on dividing the total number of sequences present by the time allowed for deposition.

Unit 3 is similar to Unit 2 with stacked stratal packages that are divided into three sub-units. Sub-unit 3a consists of single-story channel-fill sandstones separated by overbank fines. Sub-unit 3b consists of small-scale, isolated channel-fill

Figure 3 (on following page): Figure showing how accommodation changes within the three units of the Williams Fork Formation. The vertical line represents zero accommodation. To the right, the rate accommodation increases. To the left there is negative accommodation (erosion). Unit 1 shows relatively constant accommodation. Each depositional sequence in Unit 2 shows increasing accommodation from Sub-unit 2a to Sub-unit 2c. At the sequence boundary there is negative accommodation. In Unit 3 each depositional sequence shows increasing accommodation from Sub-unit 3a to Sub-unit 3b with a slight decrease in the rate of accommodation in Sub-unit 3c.



Rate of Accommodation
–
+
UNIT 3
UNIT 2
UNIT 1
Stratal Package
cs
ms
fs
vfs
Silt
Shale
KEY
Shale
Siltstone
Sandstone
Channel-Fill Sandstone Lense
SP Stratal Package
Sequence Boundary
vfs Very Fine Sand
fs Fine Sand
ms Medium Sand
cs Coarse Sand

sandstones encased in overbank fines, and Sub-unit 3c consists of paleosols.

Each stratal package in Unit 3 records a change from low accommodation at the base, an increase in accommodation and then a slight decrease in accommodation at the top represented by the paleosols (Fig. 3). Each represents a depositional sequence bounded by sequence boundaries and contains a lowstand to transgressive systems tract (Sub-unit 3a) and a highstand systems tract (Sub-units 3b and 3c) (Fig. 2). Sequence boundaries within Unit 3, as in Unit 2, are estimated to be fourth order.

ACCOMMODATION

Mechanisms controlling deposition of the Williams Fork Formation must be interpreted on two different time scales. The first represents the mechanism(s) that control the longer-term subsidence that formed accommodation throughout deposition of the formation (time scale of several million years). The second represents the mechanism(s) controlling the shorter-term (~0.4 to 0.6 ma) oscillations in base level that produced each of the depositional sequences within Units 2 and 3.

Mechanism(s) controlling longer-term changes in accommodation

The stacking of multiple sequences indicates that a minimum of 500 m of accommodation formed during deposition of the Williams Fork Formation. Accommodation can be formed by either a eustatic rise (Jervey, 1988; Shanley and McCabe, 1994) or by subsidence related to thrust sheet or sediment loading (Fleming and Jordan, 1989). While thrust-sheet loading was an active process during the Campanian and Maastrichtian in the Sevier thrust belt, this time interval was also characterized by a third order eustatic fall (Kennett, 1982; Haq et al, 1987), which excludes a eustatic rise as a mechanism to form the accommodation. This third-order fall was punctuated by smaller scale (fourth or fifth order) rises and falls producing small scale transgressions and regressions of the shoreline and the associated base level changes in the fluvial section. These fourth and fifth-order changes may have produced the multiple depositional sequences within Units 2 and 3.

Within foreland basins, tectonic subsidence produces accommodation and forms from isostatic readjustment associated with the emplacement of thrust sheets (Quinlan and Beaumont, 1984). Thrust-sheet loading results in greater subsidence (i.e. greater accommodation) immediately adjacent to the thrust load, with the magnitude of subsidence decreasing with distance away from the thrust load (Quinlan and Beaumont, 1984). As the thrust sheet erodes, the load is redistributed to the basin as sediment fill, and can account for up to 60% of the total subsidence within a foreland basin (Fleming and Jordan, 1989). The precise location of the study area within the foreland basin, with respect to the thrust belt and peripheral bulge, is uncertain. The study area is estimated to lie approximately 215 km east of the Gunnison thrust (the active thrust during the latest Cretaceous) (DeCelles, 2004; DeCelles and Coogan, 2006). The exact position of the forebulge during the latest Cretaceous is controversial but is thought to be located in central Colorado, which would place it approximately 100 to 200 km east of the study area (DeCelles and Currie, 1996; White et al., 2002). The location of the study area within the foreland basin suggests that sediment loading likely played the largest role in the formation of subsidence (i.e. accommodation) within the study area (e.g. Fleming and Jordan, 1989).

Without overall subsidence to produce accommodation, successive rises and falls in base level will continuously rework sediment in the same stratigraphic interval and will not produce vertically stacked deposits, such as those seen in the Williams Fork Formation. For example, Quaternary strata of the Mississippi River system record deposition during multiple glacio-eustatic oscillations at a time when new accommodation was not formed. These deposits are preserved as disconnected terrace remnants along valley margins and not as a thick, laterally continuous stratigraphic succession (Blum et al., 2000). The terrace remnants reflect several glacial stages but are preserved as reworked sediment in the same stratigraphic interval because the glacio-eustatic oscillations led to the continuous reworking of the same deposits. Similar conditions also occurred in the Cretaceous Dakota Group of Colorado and New Mexico where base-level fluctuations led to repeated valley-scale cut-and-fill cycles in the absence of new accommodation (Holbrook et al., 2006). In the Williams Fork Formation, if there were no net formation of accommodation, the multiple stacked depositional sequences (net accumulation of ~500 m) would not be preserved.

Mechanism(s) controlling shorter-term changes in accommodation

Stratal packages in Units 2 and 3 are interpreted to indicate a complicated interplay between subsidence and regional erosion: subsidence (producing accommodation) is required to form the space needed for the deposition of the multiple stacked depositional sequences, however, a fall in base level, resulting in regional erosion, destroys accommodation and results in the sequence boundaries bounding each depositional sequence.

Short-term falls in base level can be produced tectonic uplift (Devlin et al., 1993), eustatic fall, and changes in sediment supply (Miall, 1996) (Fig. 4). Structural movements within the foreland basin, such as blind thrusting, can produce localized uplift resulting in sequence boundaries (Bridge, 2006) (Fig. 4). The study area, however, shows no evidence for structural uplift during deposition (e.g. faulting, offset strata or thickness differences attributed to structural movement). In areas immediately adjacent to the thrust sheet, isostatic rebound, due to erosion of the thrust sheet, could result in enough uplift within the basin to form sequence boundaries (Catuneanu et al., 2000; Willis, 2000; Catuneanu and Elango, 2001) (Fig. 4). This response results in a tectonic cyclicity produced by times of active thrusting alternating with periods of tectonic quiescence that alternately form and destroy accommodation. During times of active thrusting, the resultant load increases subsidence within the adjacent basin forming accommodation. After thrusting ceases, the thrust belt erodes, thinning the thrust belt and resulting in isostatic rebound. The isostatic rebound destroys accommodation resulting in further erosion and the formation of a sequence boundary. Thick fluvial and marine intervals from the late Cretaceous to Paleocene of the Western Canada foreland basin are interpreted to record this tectonic cyclicity on a scale of 0.2 to 0.5 Ma (Catuneanu et al., 2000).

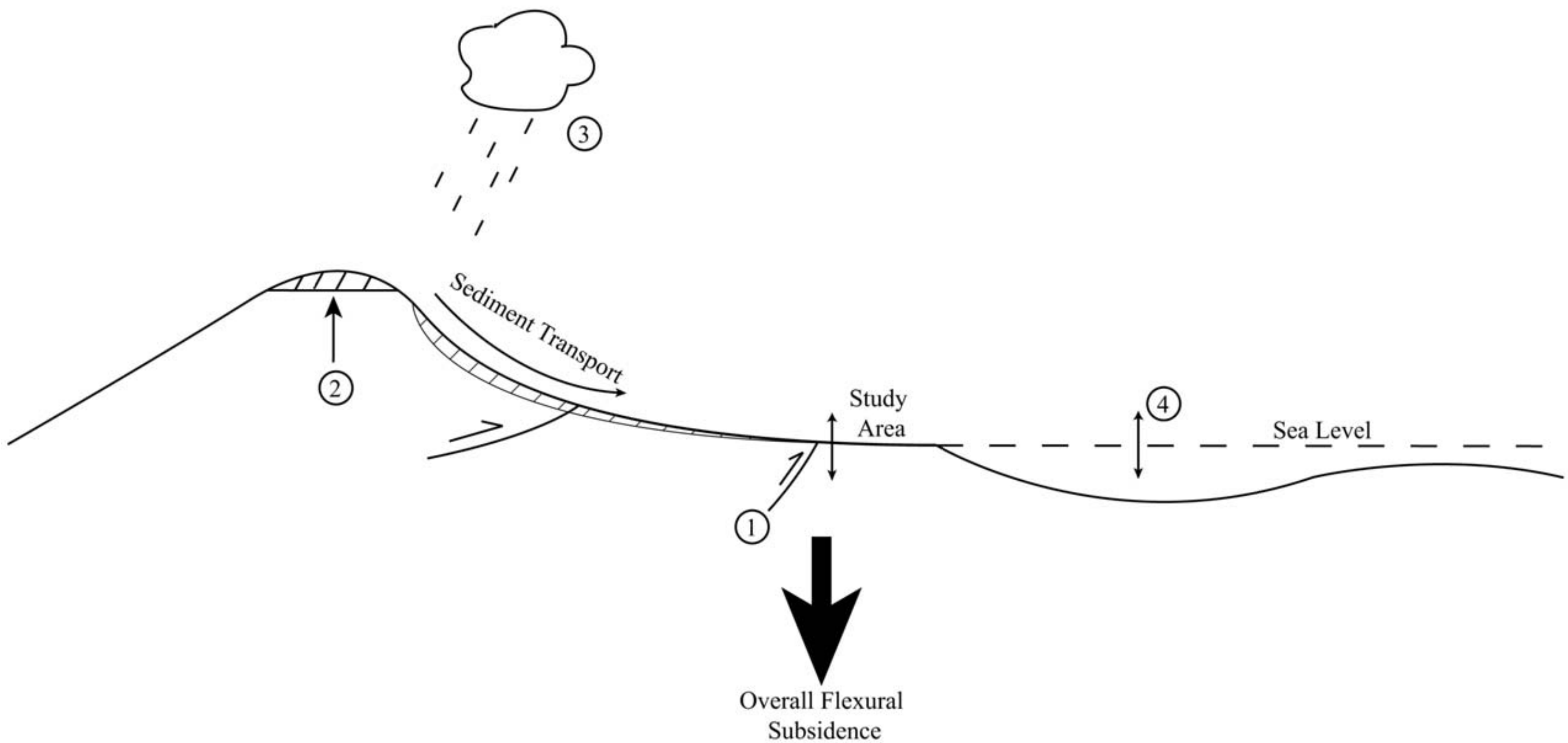


Figure 4: Schematic illustrating several mechanisms that are capable of producing shorter-term changes in accommodation within a foreland basin. Hatch marks represent areas eroded. There are four possible mechanisms that may be controlling the shorter-term changes in accommodation within Units 2 and 3 of the Williams Fork Formation. 1) Blind thrusting within the distal portions of the foreland basin can produce localized uplift. 2) Isostatic rebound within can occur as a result of erosion within the fold and thrust belt. 3) Changes in climate can result in cyclic changes in sediment supply. A decrease in sediment supply can result in erosion while an increase in sediment supply can lead to deposition within fuvial systems. Less erosion can occur in areas where the fuvial system is already at base level. 4) Oscillations in eustasy can result in shorter-term oscillations in base level within the study area.

Isostatic rebound, however, affects accommodation in regions proximal to the orogenic belt (Catuneanu and Elango, 2001) whereas the Williams Fork Formation lies in the distal part of the foreland basin. Also, sediment eroded off of the thrust belt during isostatic rebound is then distributed within the foreland basin resulting in increased subsidence due to sediment loading (Fleming and Jordan, 1989), which can reduce the likelihood that sequence boundaries will form. The distance of the study area from the orogenic belt and the subsidence resulting from sediment loading eliminates isostatic rebound as a viable mechanism for producing sequence boundaries within the Williams Fork Formation.

Changes in sediment supply are capable of producing sequence boundaries within fluvial systems. If sediment supply is reduced then the fluvial system can erode, producing a sequence boundary (Miall, 1996) (fig. 4). If the sediment supply is increased then the fluvial system will aggrade (Miall, 1996). Climatic changes are capable of producing changes in sediment supply on multiple time scales by controlling the amount of erosion (Miall, 1996). There is no evidence of dramatic shifts in climate within the study area during deposition of the Williams Fork Formation. Also, fluvial systems cannot erode below base level. If a fluvial system has a very low gradient and is near base level, then a reduction in sediment supply will not lead to formation of a sequence boundary because the fluvial system has no room to erode. Unit 1 and sub-unit 2c are both interpreted to be low gradient fluvial systems. Unit1 and each occurrence of sub-unit 2c are capped by sequence boundaries and overlain by higher energy systems representing a higher gradient fluvial system. A reduction in sediment supply after deposition of Unit 1 or sub-unit 2c would have led to erosion but could have formed the sequence boundaries resulted in the overlying higher gradient fluvial systems because the systems were already near base level and had little room to erode any further.

Eustatic fall is a potential mechanism to form the sequence boundaries bracketing the depositional sequences within in the Williams Fork Formation (Fig. 4). Sea-level oscillations can affect the gradient of fluvial systems a significant distance up depositional dip. Studies of Quaternary fluvial systems, such as the Mississippi River, demonstrate that eustatic changes influence fluvial systems as far inland as 200 to 300 km (Aslan and Autin, 1999). Studies of the Pleistocene and Holocene deposits of the Colorado River in Texas indicate fluvial incision caused by eustatic fall. Approximately 35 m of fluvial incision was documented at the coeval shoreline and approximately 15 to 20 m of fluvial incision was documented 100 to 150 km up depositional dip of the coeval shoreline (Blum and Price, 1998). The incision described results from drops in base level (sea level) corresponding to 100,000 yr glacioeustatic oscillations.

While many believe that the Cretaceous was free of ice caps (Kennett, 1982; Huber et al., 2002), recent studies have proposed restricted development of ice sheets during this time (Miller et al., 2003; Crampton et al., 2006). Rapid (< 1 million years) eustatic falls during the latest Cretaceous, greater than 25 m, are proposed to represent a glacio-eustatic control caused by the development of small, ephemeral ice sheets in Antarctica. Data concerning fourth order eustatic fluctuations during the

Late Cretaceous is lacking, however, global sea-level charts indicate that fourth-order changes occur even during periods known to be free of ice caps (Haq et al., 1987; Haq and Schutter, 2008). The fourth-order cycles represent 0.2 to 0.5 Ma and are a magnitude of 30 m or less (Vail et al., 1977).

UNIT 1

Unit 1 records the continuous formation of accommodation through time. The parasequences within the unit are aggradationally stacked, and channel-fill sandstone and overbank characteristics (i.e. grain size, cross-bed thickness, color) are consistent among parasequences. The consistency of the facies throughout Unit 1 suggests that the gradient of the fluvial system remained relatively unchanged during deposition, and that the rate of basin subsidence was equal to the rate of sediment influx, resulting in the aggradationally stacked parasequences. If a relative rise or fall in base level had occurred, a change in fluvial facies reflecting an increase or decrease in fluvial energy would be expected.

Sequence boundaries are common within deposits of the Western Interior foreland basin, and it is unusual for sequence boundaries to be missing in a succession as thick as Unit 1 (~120 m). The Mt. Garfield Formation, which underlies the Williams Fork Formation, and the upper two units of the Williams Fork Formation both contain sequence boundaries at a high frequency interpreted to reflect multiple eustatic falls (Zater, 2005; Madof, 2006). The timing of deposition of the Williams Fork Formation is disputed, making it difficult to match eustatic curves with Units 2 and 3 (Johnson and May, 1980; Patterson et al., 2003). It is unlikely, however, that eustatic fluctuations occurred during deposition of older and younger units but not during deposition of Unit 1. If the rate of subsidence during deposition of Unit 1 is higher than the rate of eustatic fall, the effect of any eustatic fall would be a relative rise in base level, and sequence boundaries would not be produced. While this can explain the absence of identified sequence boundaries, the presence of eustatic oscillations should still produce periods of higher and lower accommodation. The rate of accommodation will increase as a eustatic rise adds to the accommodation formed by subsidence and the rate of accommodation will decrease as a eustatic fall offsets part of the accommodation formed by subsidence. These repetitive changes in accommodation should result in sections of less channel-fill connectivity reflecting the periods of higher accommodation and sections of more channel-fill connectivity reflecting the periods of lower accommodation. No such pattern is identified within Unit 1. It is possible that these stratal patterns may be present but are poorly developed and difficult to identify in the field due to poor exposure of the slope-forming Unit 1. A recent study combining measured stratigraphic sections with high-resolution aerial light detection and ranging (LIDAR) data resulted in high-resolution images of sandstone connectivity in the lower Williams Fork Formation (Pranter et al., 2009). While the purpose of the study was not to identify stratal packaging, cursory observations of the data available suggest that there may be subtle stratal packages present within Unit 1. More work with the LIDAR data is needed to confirm and describe these stratal packages.

UNIT 2

Unit 2 contains sixteen depositional sequences, each bounded by a surface of regional erosion (sequence boundary), indicating a drop in base level. Stratal fill overlying each sequence boundary reflects a complicated rise in base level. The interplay between the short-term eustatic cycles and long-term tectonic subsidence may control deposition of the stacked depositional sequences in Unit 2. Short-term eustatic oscillations may control deposition of each individual depositional sequence, while the long-term tectonic subsidence may control the overall formation of accommodation. Each depositional sequence within Unit 2 is proposed to record one such short-term eustatic cycle. A single depositional sequence is proposed to begin with a fall in base level (eustatic fall). This fall leads to an increased gradient in the fluvial system resulting in regional erosion and the formation of the basal sequence boundary. This is followed by a slow rise in sea level, resulting in deposition of high-energy fluvial deposits of the late lowstand systems tract (Sub-unit 2a). The channel-fill deposits of the late lowstand are nested indicating a steady, continuous sea-level rise (Van Wagoner, 1995; Richards, 1996).

Laterally continuous channel-fill sandstones, separated vertically by overbank fines (Sub-unit 2b), overlie the nested channel-fill sandstones and are interpreted as transgressive systems tract deposits. This change is in response to an increase in the rate of base-level (sea-level) rise resulting in an increase in accommodation. Each channel-fill sandstone and associated overbank deposit represents a well-developed parasequence. Each vertically stacked parasequence is successively thinner and consists of a smaller average grain size than that of the underlying parasequence. These parasequences display a backstepping parasequence stacking pattern indicating that the rate of base-level (sea-level) rise outpaces sediment supply. The presence of well-developed parasequences indicates overall base-level (sea-level) rise was punctuated with still stands during deposition of the transgressive systems tract. Constant and continuous sea-level rise produces aggradational and nested channel-fills such as those in the late lowstand.

The transgressive systems tract is overlain by the highstand systems tract deposits that consist of isolated channel-fill sandstones encased in overbank fines (sub-unit 2c). The top of each channel-fill deposit represents the top of a parasequence. Parasequences are aggradationally stacked indicating that the rate of base-level (sea-level) rise is equal to the rate of sediment influx. This is consistent with a reduction in the rate of base-level (sea-level) rise during the highstand systems tract. The depositional sequence is truncated by a sequence boundary indicating a drop in base level (sea-level) and is in turn overlain by a new depositional sequence.

For sequence boundaries to form, the rate of eustatic fall must be greater than the rate of tectonic subsidence. The rate of subsidence from thrust loading decreases with distance from the orogenic belt (Quinlan and Beaumont, 1984) while the rate of eustatic change is constant at every point within the basin. Where and if the rate of subsidence is equal to the rate of eustatic fall, an equilibrium point is established (Jervey, 1988). Toward the thrust belt from the equilibrium point, a eustatic fall can never outpace the rate of subsidence and therefore sequence boundaries will not form

(Jervey, 1988). Basinward (toward the forebuldge) from this point, a eustatic fall, which outpaces subsidence, results in the formation of a sequence boundary. The location of the equilibrium point will change through time with changes in the rate of eustatic fall or with changes in rate of subsidence within the foreland basin (Posamentier and Allen, 1993). The presence of sequence boundaries in Unit 2 indicates that the equilibrium point was west of the study area (toward the thrust belt) resulting in eustatic falls capable of outpacing subsidence.

UNIT 3

Interplay between tectonic subsidence and eustatic cyclicity also controls deposition of the three depositional sequences in Unit 3. Depositional sequences of Unit 3 differ from those of Unit 2. In Unit 3, nested sandstone complexes representing the lowstand systems tract are absent and instead laterally continuous, channel-fill sandstones interbedded with thin deposits of overbank fines overlie the sequence boundary (Sub-unit 3a). Locally, however, the channel-fill sandstones are vertically amalgamated. Vertically, each successive channel-fill sandstone becomes progressively thinner and has a smaller average grain size (from fine upper to very fine upper.) Each channel-fill sandstone and its laterally equivalent overbank fines, represents a parasequence. The upward, progressive decrease in grain size represents an upward decrease in channel gradient (Richards, 1996) and is interpreted to represent a backstepping parasequence stacking pattern within a transgressive systems tract. As in Unit 2, the well-developed parasequences of the transgressive systems tract imply that sea-level rise was punctuated and not continuous and steady.

The transgressive systems tract is overlain by isolated channel-fill sandstones encased in overbank fines (Sub-unit 3b), which is then overlain by a well-developed paleosol (Sub-unit 3c). This succession represents an overall decrease in the rate of base-level (sea-level) rise and is interpreted as deposition during the highstand systems tract. Parasequences are aggradationally stacked.

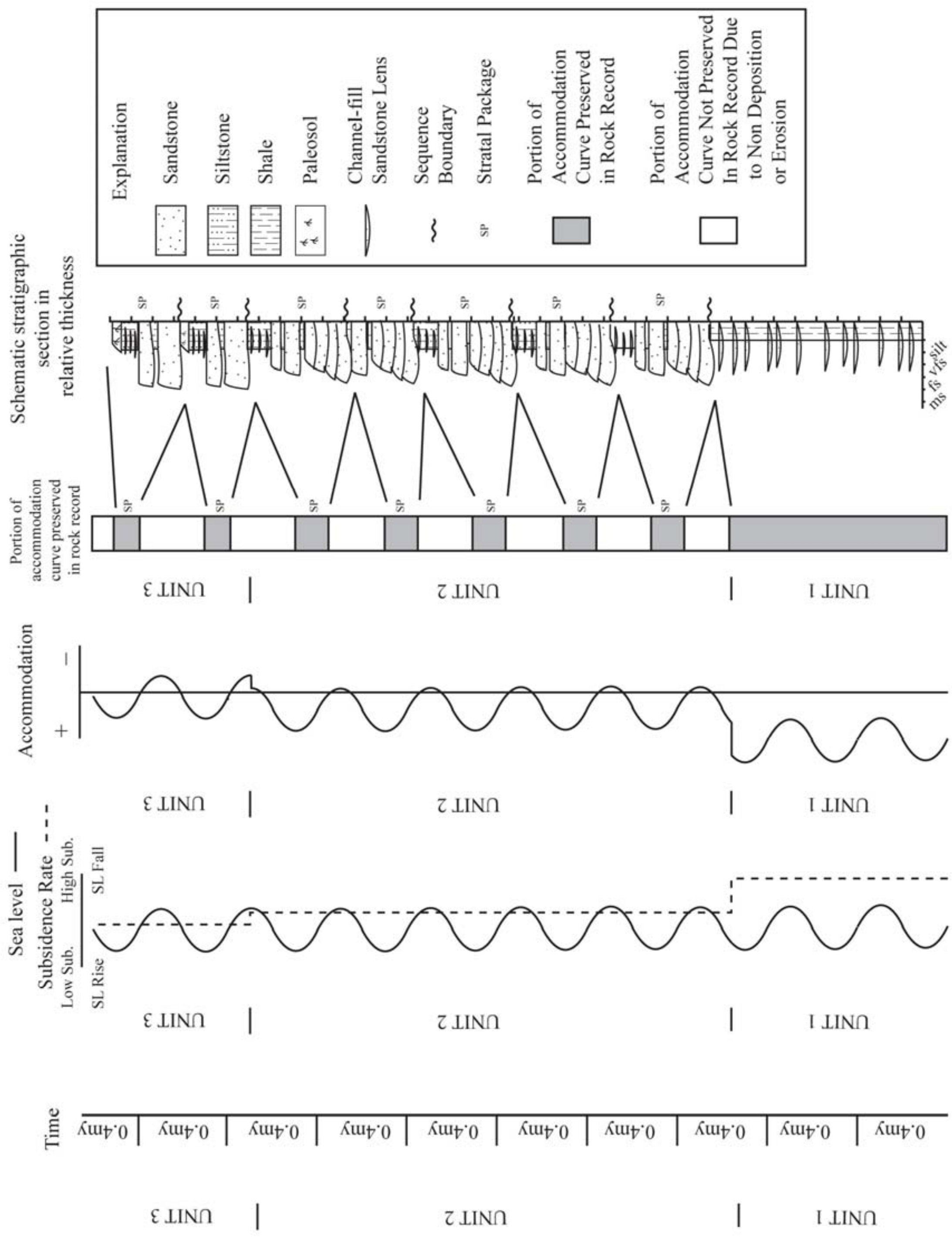
There are two possible explanations for the absence of well-developed lowstand systems tract deposits in Unit 3. One possibility is that deposits in Unit 3 accumulated farther from the coeval shoreline than the deposits in Unit 2 and that the well-developed lowstand deposits occur down dip of the coeval transgressive systems tract deposits of Unit 3. This interpretation is used to explain the lack of lowstand systems tract deposits overlying sequence boundaries in late Cretaceous fluvial deposits of the Kaiparowits Plateau (Shanley and McCabe, 1995). During a base-level rise, fluvial systems fill basinward resulting in aggradation of the down-dip portions of a fluvial profile while up-dip portions may still be eroding or in a steady state condition (Richards, 1996). In the up-dip sections, there may not be a time of slow base-level rise. A slow base-level rise is needed for deposition of a nested channel-fill complex, typical of the lowstand systems tract. Another explanation is that the lowstand systems tract is not well-developed in Unit 3 because sea-level oscillations during deposition of Unit 3 are more rapid than those during deposition of Unit 2. A faster rise in sea level would explain the absence of nested channel-fill deposits of the lowstand interpreted to be deposited during a time of slow, continuous sea-level rise.

CHANGES IN ACCOMMODATION STYLE THROUGH TIME

Overall the changes in stratal packaging indicate a change in accommodation style from Unit 1 to Unit 3: this is seen through an overall decrease in the preservation of overbank deposits, an increase in the frequency of sequence boundaries and a decrease in the average thickness of depositional sequences. The stratal packaging, and consequently the accommodation style, is controlled by the interplay between eustatic fluctuation and tectonic subsidence (Fig. 4). The overall changes in accommodation style observed between the three units are controlled by changes in the rate of tectonic subsidence or changes in the magnitude and rate of eustatic fluctuations.

The first major change in stratal packaging is between Unit 1 and Unit 2. Unit 1 has no identifiable sequence boundaries and its stratal packaging indicates a high rate of accommodation. Unit 2 contains multiple depositional sequences, each bounded by sequence boundaries, which indicate that accommodation was alternately formed (allowing for deposition of the depositional sequences) and then destroyed (forming the sequence boundaries). There are two mechanisms that may cause this change in accommodation style. One possibility is that tectonic subsidence is constant between Units 1 and 2 but eustatic fluctuations are absent during deposition of Unit 1 and present during deposition of Unit 2. As previously discussed, however, it is likely that eustatic oscillations did occur during deposition of Unit 1.

Figure 5 (On following page): Schematic displaying how sea level and subsidence interact to form the rock record preserved. All rates are relative and non-quanitative. The first column shows equal units of time (0.4 Ma). The second column shows sea-level oscillations through time and changes in the relative rate of subsidence through time. Each sea-level oscillation occurs in approximately 0.4my. This timing is an approximation based on dividing the total time represented by the Williams Fork Formation by the number of sequences. Sea-level oscillations remain relatively constant throughout deposition of each of the three units while the rate of subsidence decreases from Unit 1 to Unit 2 and again from Unit 2 to Unit 3. In Unit 1 the rate of sea-level fall never outpaces the rate of subsidence while in Units 2 and 3 the rate of sea-level fall does outpace the rate of subsidence at times. The third column shows total accommodation through time as a combination of the effects of sea level and subsidence. During deposition of Unit 1 there was always accommodation. Units 2 and 3 show periods of formation of accommodation and periods of destruction of accommodation. Column four shows periods of time that are preserved in the rock record and periods that are not. Notice all of Unit 1 is preserved while only portions of Units 2 and 3 are preserved. This is due the erosion of the rock record at times when there is negative accommodation. The only portion of the accommodation curve that is preserved in the rock record is the increase in accommodation from the base of each stratal package to the top. The final column shows how the stratigraphic section, in thickness, relates to time. The periods of erosion that are no longer preserved appear only as sequence boundaries in the rock record while the periods of deposition form the stratal packages identified within Units 2 and 3. Figure does not represent total amount of time represented by the Williams Fork Formation.



Time
0.4my
Sea level
Subsidence Rate
Low Sub.
High Sub.
SL Rise
SL Fall
Accommodation
+
−
UNIT 1
UNIT 2
UNIT 3
Portion of accommodation curve preserved in rock record
SP
Schematic stratigraphic section in relative thickness
silt
vfs
fs
ms
Explanation
Sandstone
Siltstone
Shale
Paleosol
Channel-fill Sandstone Lens
Sequence Boundary
Stratal Package
Portion of Accommodation Curve Preserved in Rock Record
Portion of Accommodation Curve Not Preserved In Rock Record Due to Non Deposition or Erosion

The second possibility is that the rate and magnitude of eustatic fluctuations remained relatively constant between Units 1 and 2, but the rate of tectonic subsidence was higher during deposition of Unit 1. A subsidence rate that is rapid enough to outpace the rate of eustatic fall could explain the lack of sequence boundaries in Unit 1. During deposition of Unit 2, if the subsidence rate was less than the rate of eustatic fall, sequence boundaries could form. Although the timing of deposition of the Williams Fork Formation is disputed, average subsidence rates can be calculated for the end-member estimates of the duration of deposition. If the Williams Fork Formation represents 8 Ma of deposition then the average subsidence rate is ~3.5 cm/1000 yr. If the Williams Fork Formation represents 14 Ma of deposition then the average subsidence rate is 6.5 cm/1000 yr. These rates are derived by dividing the total thickness of the Williams Fork Formation in the study area by the time allowed for deposition. These rates do not take into account the uncompacted sediment thickness and therefore represent the lower end members for subsidence rates. These rates are similar to subsidence rates calculated for the distal portion of the Western Interior foreland basin at various times and locations (~ 3.3 to 10 cm/1000 yr) (Devlin et al., 1993; Willis, 2000; DeCelles, 2004). The rate of eustatic fall must be lower than the subsidence rate in Unit 1 in order to explain the lack of identified sequence boundaries. In Unit 2, the rate of subsidence must at times outpace the rate of any eustatic falls (during deposition of the depositional sequences), and at times the rate of eustatic fall must outpace the rate of subsidence (in order to form sequence boundaries). The rate of eustatic fall, due to proposed glacio-eustatic controls at the end of the Cretaceous, are estimated to be as high as 4 cm/1000 yr with the magnitude of eustatic fall averaging 25 m (Miller et al., 2003). The subsidence rates calculated are average rates for the entire formation resulting in some periods of higher than average subsidence and other periods of lower than average subsidence. The absence of identified sequence boundaries in Unit 1 suggests that it was deposited during a time of higher than average accommodation, and if subsidence rates were greater than the rate of any eustatic falls (up to 4 cm/1000 yr), sequence boundaries would not have formed. During deposition of Unit 2, if the average rate of subsidence was lower and if, at times, the rate of eustatic fall outpaced the rate of subsidence, sequence boundaries would have formed.

The second overall change in accommodation style is observed between Units 2 and 3. Depositional sequences in Unit 2 are thicker, on average, than the depositional sequences in Unit 3 (~ 25 m in Unit 2 and ~16 m in Unit 3). This is interpreted to represent a reduction in the rate of accommodation from Unit 2 to Unit 3. This reduction in the rate of accommodation may represent a reduction in the rate of subsidence. If eustatic fluctuations were similar during deposition of Units 2 and 3, a reduction in the rate of subsidence would result in an increase in the rate in which sequence boundaries formed, resulting in thinner depositional sequences in Unit 3 than Unit 2.

Changes in the structural setting of the Cretaceous Western Interior foreland basin may explain the proposed decrease in subsidence rate throughout deposition of the formation. The Williams Fork Formation is the last formation deposited within the Cretaceous Western Interior foreland basin of Colorado. The boundary between the Cretaceous Williams Fork Formation and the overlying Paleocene Wasatch Formation represents a change in tectonic processes from thin-skinned deformation to thick-skinned deformation (DeCelles and Coogan, 2006). This change may have occurred gradually during deposition of the Williams Fork Formation. As thin-skinned deformation ceases, the rate of tectonic subsidence from thrust emplacement or both the rate of thrust emplacement and sediment loading began to decrease. The decrease in the rate of tectonic subsidence may account for the overall decrease in accommodation throughout deposition of Williams Fork Formation.

CONCLUSIONS

The stratal patterns seen within the Williams Fork Formation are proposed to record an interaction between eustatic fluctuations and tectonic subsidence. Eustatic changes are proposed to control the cyclicity in base level, which formed the depositional sequences. These cyclic changes controlled the stratigraphic arrangement at a higher frequency (Sub-units 2abc, 3abc) within the large-scale stratal packages (Units 1, 2 and 3). Tectonic subsidence increases accommodation, allowing space for multiple depositional sequences to be preserved. Without continued subsidence the stratigraphy would be similar to what is described in the Cretaceous Dakota Group of Colorado and New Mexico where sea-level fluctuations led to repeated erosion and deposition within the same stratigraphic interval (Holbrook, 2006). Without net accommodation the same erosional space is filled repeatedly by reworking of the same stratigraphic interval during multiple base-level oscillations. Tectonic subsidence also controls the large scale changes in stratal packaging, which in turn reflect changes in accommodation through time. Overall, accommodation decreases from Unit 1 to Unit 3 suggesting a decrease in the rate of subsidence through time. This is recorded by the change from isolated channel-fill sandstones with limited interconnectivity and abundant overbank fines in Unit 1 to channel-fill sandstones with a much higher interconnectivity in Units 2 and 3. The decrease in the thickness of depositional sequences between Units 2 and 3 also records a reduction in accommodation. This reduction in accommodation may be due to the cessation of thrust-related subsidence during the final stages of the Western Interior foreland basin as the change from thin-skinned to thick-skinned tectonics occurs.

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APPENDIX I

Composite Measured Stratigraphic Section

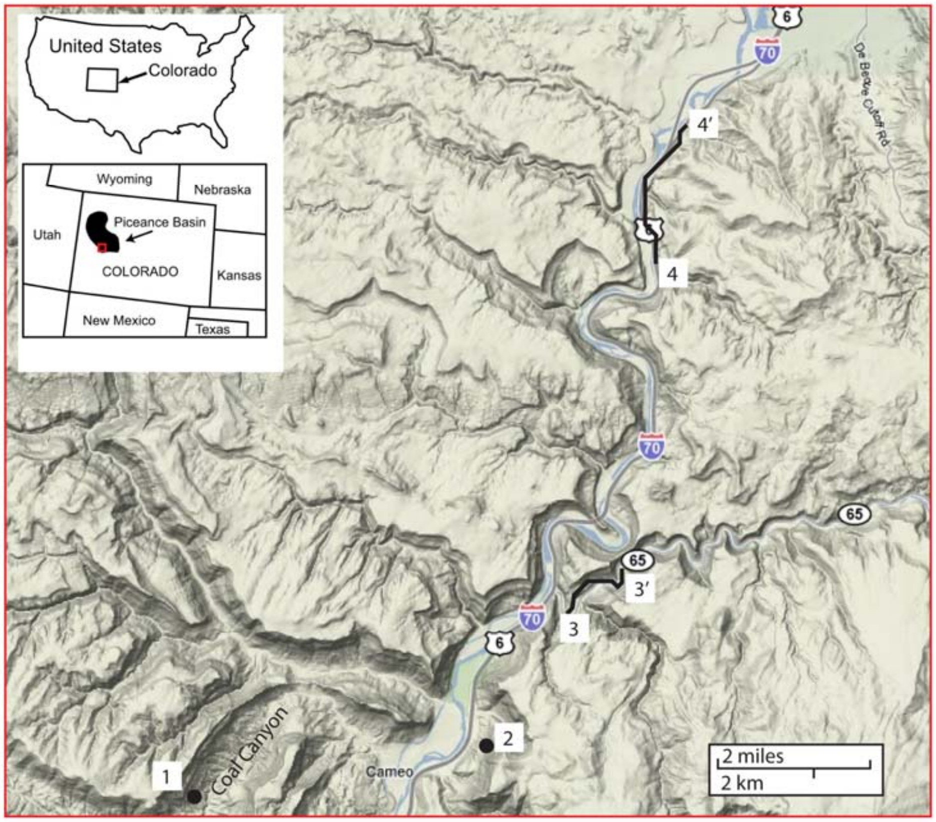
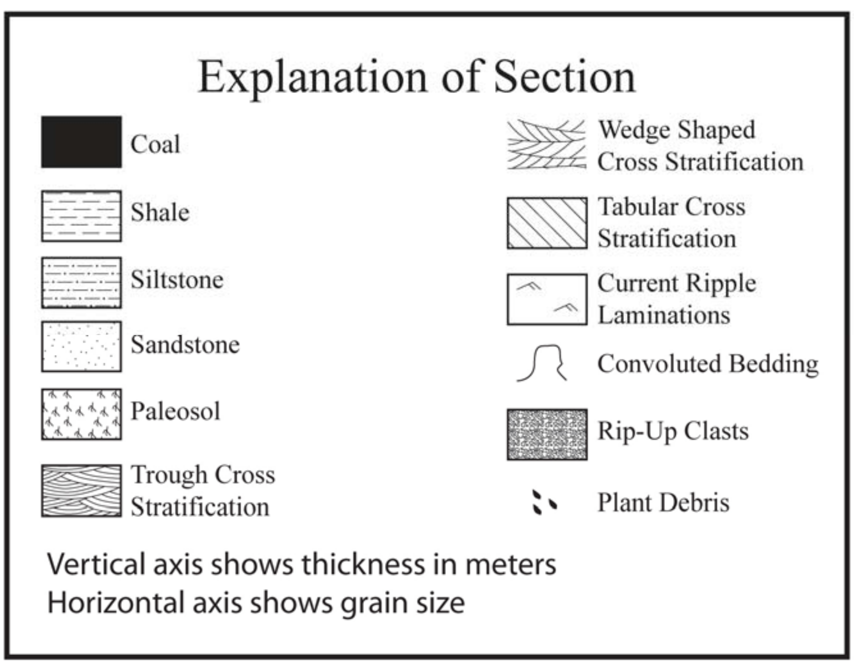
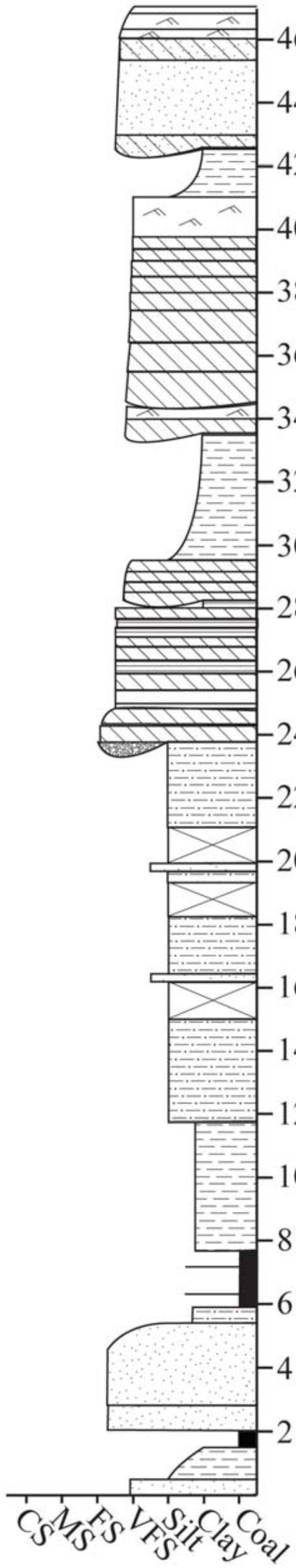


Figure 1: Map of study are showing locations of the 4 stratigraphic sections that comprise the complete stratigraphic section. Sections 1 and 2 are vertical sections measured in one place. Sections 3-3´ and 4-4´ are vertical sections that were measured along a transect. (Figure modified from Google Maps)

Table 1: Table showing which portion of the composite stratigraphic sections each individual stratigraphic section comprises. Table also shows locations of stratigraphic sections in latitude and longitude as well as township and range.

Section	Position within composite section	Latitude and Longitude	Township and Range
1	0m to 98m	39°7′59′′N 108°22′14′′W	SE1/4 of NW1/4 of Section 5, T11S R98W
2	98m to 130m	39°09′2.0′′N 108°17′49′′W	SE1/4 of SW1/4 of Section 26, T10S R98W
3-3’	130m to 246m	39°10′52′′N 108°16′26′′W to 39°11′16′′N 108°15′51′′W	SW1/4 of SE1/4 of Section 13, T10s R97W to SE1/4 of SW1/4 of Section 18 T10S R97W
4-4’	246m to 522m	39°14′50′′N 108°15′23′′W to 39°16′21′′N 108°14′49′′W	NE1/4 of NW1/4 of Section 30 T09S R97W to NW1/4 of SW1/4 of Section 20 T09S R97W





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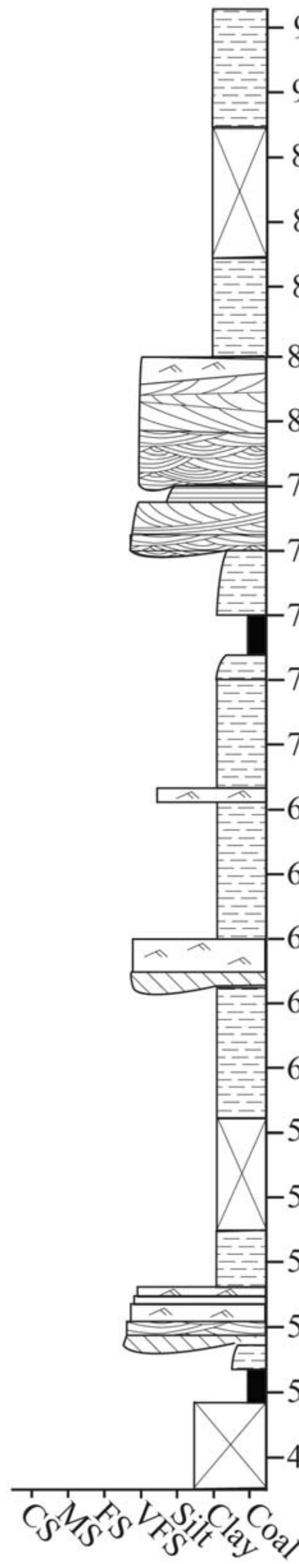
UNIT 1



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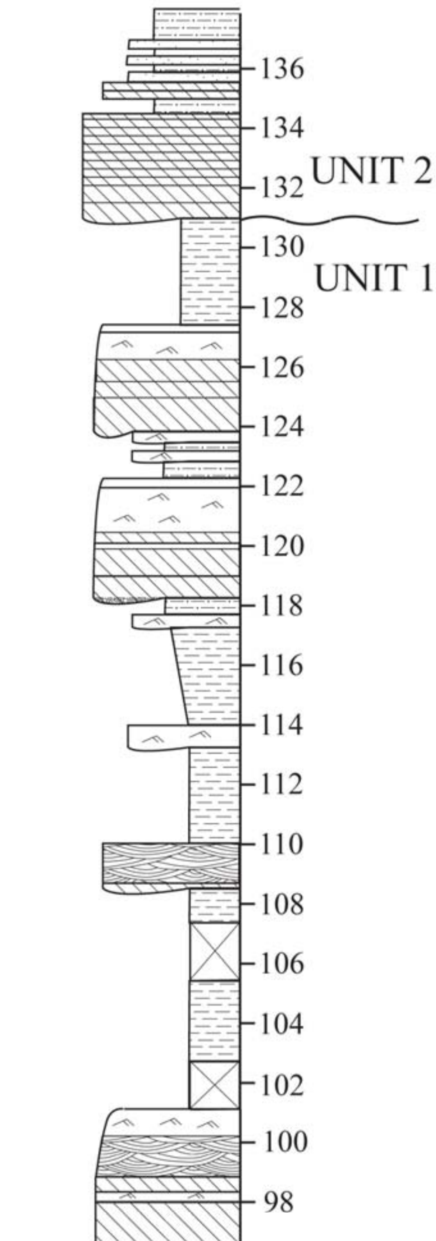
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UNIT 2
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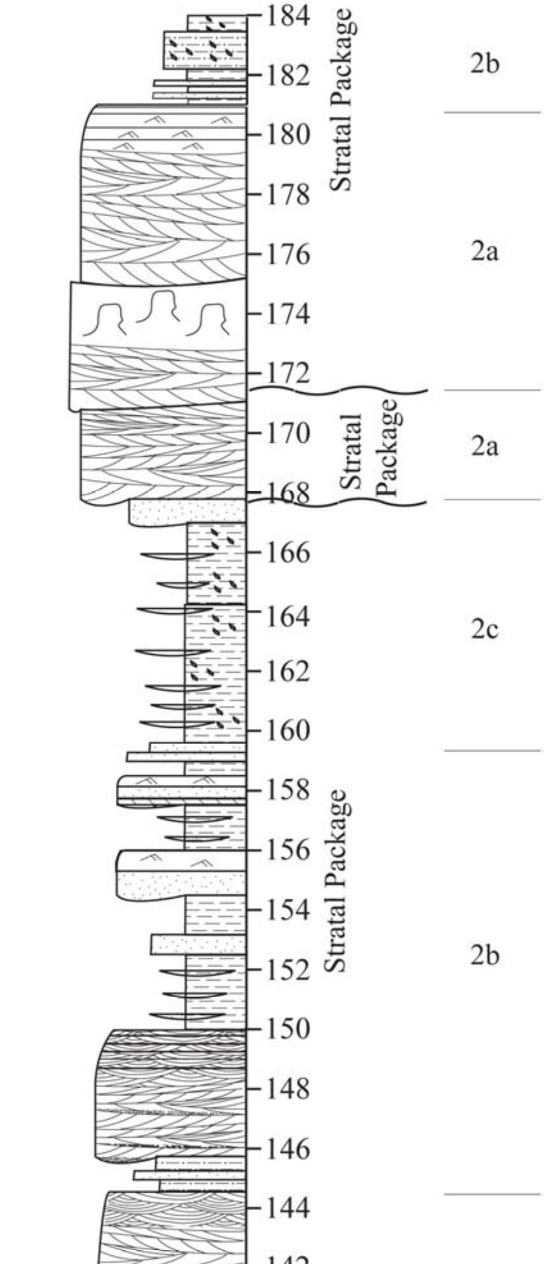
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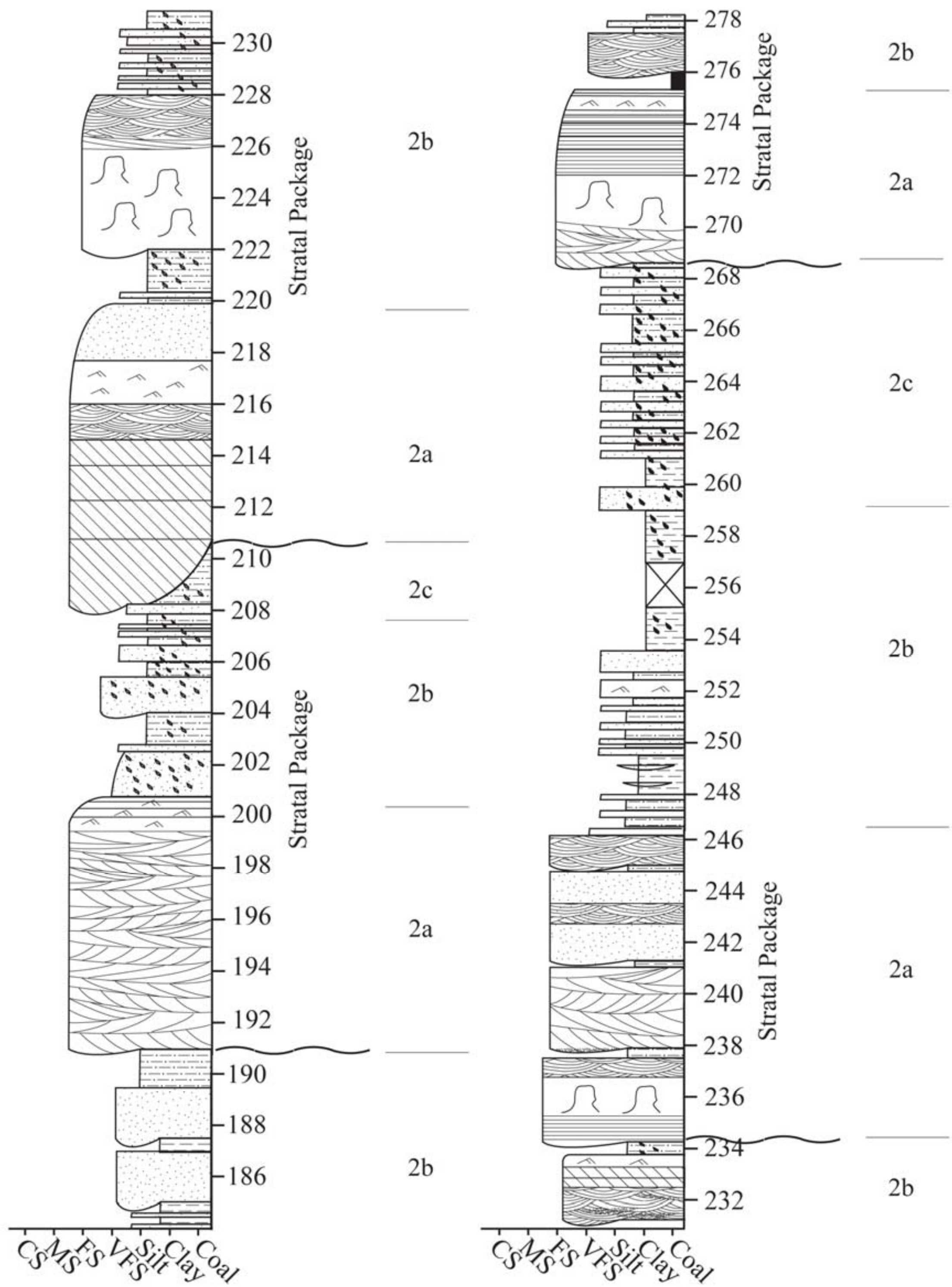

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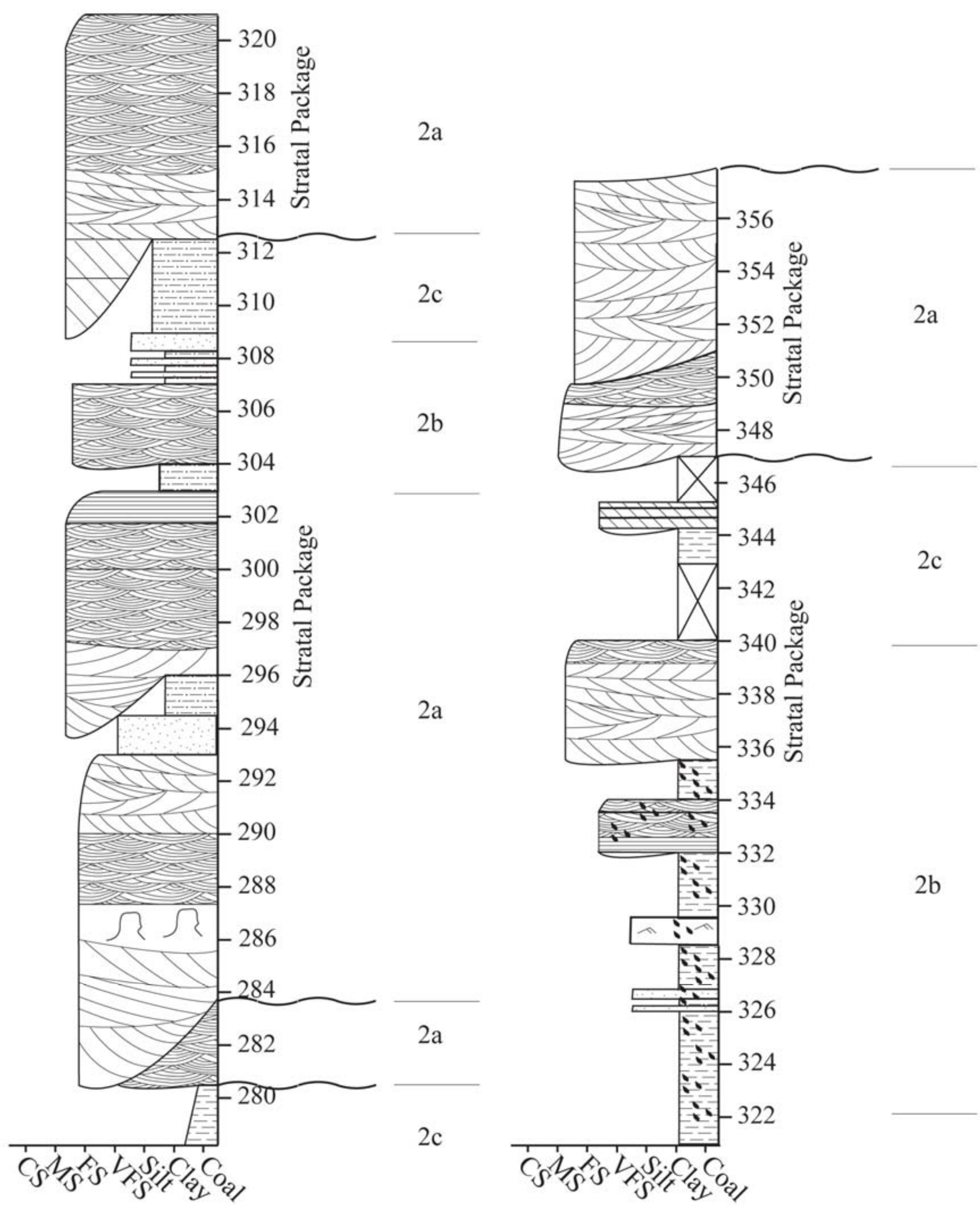

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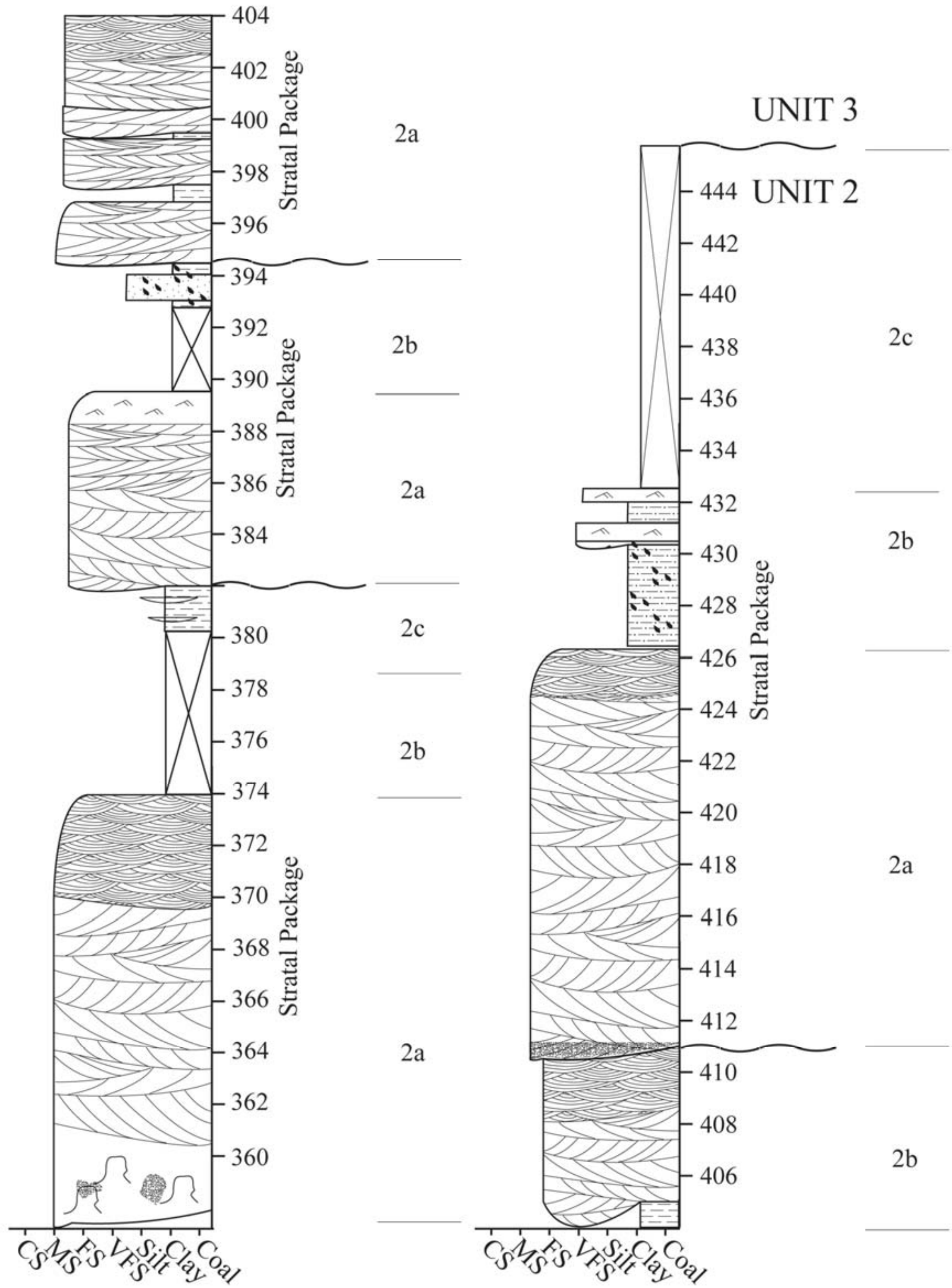
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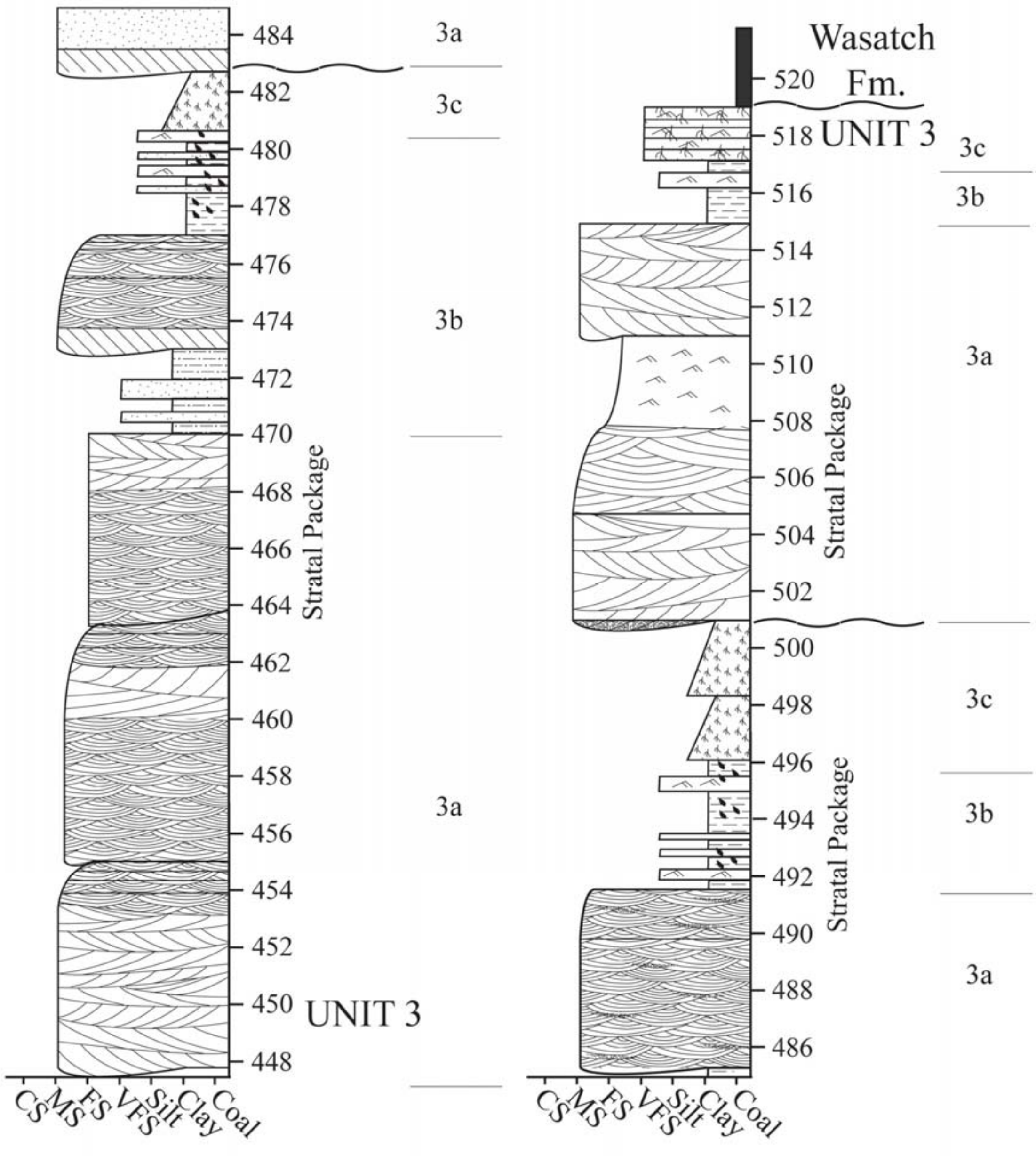

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Stratal Package
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UNIT 3
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Stratal Package
Stratal Package

References

Map of Cameo, Colorado, 2010, using Google Maps: http://maps.google.com/maps ?f=q&source=s_q&hl=en&geocode=&q=Grand+junction+Colorado&sll= 39.753657,-85.984497&sspn=0.663033,1.312866&ie=UTF8&hq=&hnear= Grand+Junction,+Mesa,+Colorado&ll=39.213635,108.303566&spn= 0.167051,0.328217&t=p&z=12, 18 January, 210